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ECONOMIC GEOLOGY

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BULLETIN OF THE SOCIETY OF
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VOL. 49 MAY, 1954 NO. 3

SEDIMENTARY FACIES OF IRON-FORMATION

HAROLD L. JAMES

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ABSTRACT

The sedimentary iron-formations of Precambrian age in the Lake Superior region can be divided on the basis of the dominant original iron mineral into four principal facies: sulfide, carbonate, oxide, and silicate. As chemical sediments, these rocks reflect certain aspects of the chemistry of the depositional environments. The major control, at least for the sulfide, carbonate, and oxide types, probably was the oxidation potential. The evidence indicates that deposition took place in restricted basins, which were separated from the open sea by thresholds that inhibited free circulation and permitted development of abnormalities in oxidation potential and water composition.

The sporadic distribution of metamorphism and of later oxidation permits description of the primary facies on the basis of unoxidized, essentially unmetamorphosed material. The *sulfide facies* is represented by black slates in which pyrite may make up as much as 40 percent of the rock. The free-carbon content of these rocks typically ranges from 5 to 15 percent, indicating that ultra-stagnant conditions prevailed during deposition. Locally, the pyritic rocks contain layers of iron-rich carbonate. The *carbonate facies* consists, in its purer form, of interbedded iron-rich carbonate and chert. It is a product of an environment in which oxygen concentration was sufficiently high to destroy most of the organic material but not high enough to permit formation of ferric compounds. The *oxide facies* is found as two principal types, one characterized by magnetite and the other by hematite. Both minerals appear to be of primary origin. The magnetite-banded rock is one of the dominant lithologies in the region; it consists typically of magnetite interlayered with chert, carbonate, or iron silicate, or combinations of the three. Its mineralogy and association suggest origin under weakly oxidizing to moderately reducing conditions, but the mode of precipitation of magnetite is not clearly understood. The hematite-banded rocks consist of finely crystalline hematite interlayered with chert or jasper. Oolitic structure is common. This facies doubtless accumulated in a strongly oxidizing, probably near-shore, environment similar to that in which younger hematitic ironstones such as the Clinton oolite were deposited. The *silicate facies* contains one or more of the hydrous ferrous silicates (greenalite, minnesotaite, stilpnomelane, chlorite) as a major constituent. Granule

structure, similar to that of glauconite, is typical of some varieties; others are nongranular and finely laminated. The most common association of the silicate rocks is with either carbonate- or magnetite-bearing rocks, which suggests that the optimum conditions for deposition ranged from slightly oxidizing to slightly reducing.

The relationship between the iron-rich rocks and volcanism, stressed by many authors, is considered by the writer to be structural, not chemical: in the Lake Superior region both iron-deposition and volcanism are believed to be related to geosynclinal development during Huronian time. In Michigan, the lower Huronian rocks are iron-poor quartzite and dolomite—typical "stable-shelf" deposits; much of the upper Huronian consists of iron-poor graywacke and slate with associated volcanic rocks —a typical "geosynclinal" assemblage. Thus the iron-rich beds of the middle Huronian and lower part of the upper Huronian were deposited during a transitional stage in structural history. The major environmental requirement for deposition of iron-formation is the closed or restricted basin; this requirement coincides in time with what would be a normal stage in evolution of the geosyncline: namely, structural development of offshore buckles or swells that subsequently develop into island arcs characterized by volcanism.

INTRODUCTION

THE term "iron-bearing formation"—or more commonly "iron-formation"—is applied in the Lake Superior region to a wide variety of rocks in which the only common denominator is a high content of primary (sedimentary) iron. In both written discussions and field mapping, the all-inclusive nature of the term has tended to mask the real differences that are present. Actually, iron-formation is a term for a large class of rocks of strongly differing original characteristics and degrees of metamorphism. Widespread and deep oxidation increases further the range in possible character of the rock. These secondary processes—metamorphism and oxidation—greatly complicate the task of determining primary lithologies; yet knowledge of original material is essential to the construction of rational patterns of distribution that, in turn, are required for intelligent exploration and development of the iron-rich beds.

According to the classic view, developed by Van Hise, Leith, and others who took part in the first comprehensive study of the region by the U. S. Geological Survey, the primary iron-formation consisted chiefly of chemically precipitated iron-carbonate and chert, with abundant greenalite (an iron silicate similar to glauconite) in some areas. This view is stated in its most complete form in Monograph 52 (Van Hise and Leith, 95).[1] Volcanism was considered to play an important part in the origin of the rocks, by contributing iron to sea water through reaction between submarine lavas and sea water and by direct contribution through gaseous and liquid emanations.

In a recent report (50) the writer suggested that the primary iron-rich rocks can be divided into four major sedimentary facies—sulfide, carbonate, silicate, and oxide—depending upon the dominant original iron mineral in the sediment. In this paper these facies of iron-formation are further described, and the relationship of the iron-rich rocks to volcanism and geosynclinal de-

[1] Numbers in parentheses refer to Bibliography at end of paper.

velopment is discussed. It must be emphasized that definition of iron-formation facies is in terms of original sedimentary features, insofar as these features can be determined; every attempt has been made to separate primary characteristics from characteristics that are due to later metamorphism or oxidation. Many of the iron-rich rocks now composed dominantly of silicates doubtless owe their origin to metamorphic reactions such as described by Tyler (91), and, of course, much of the rock characterized by iron oxides is due to later oxidation. But recognition of these processes and products in no way negates the proposition that iron-rich rocks characterized by silicates and by oxides also were primary sedimentary types; the later processes simply have increased

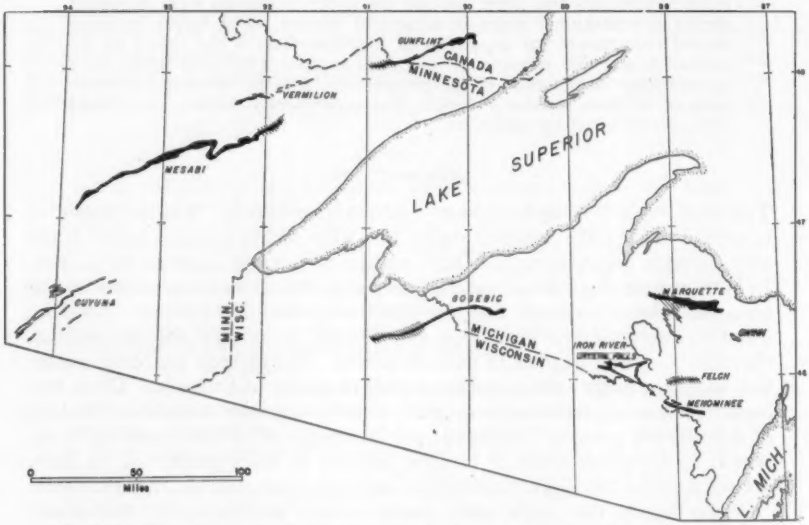

FIG. 1. Distribution of the major iron districts of the Lake Superior region. Areas of moderate- to high-grade metamorphism indicated by cross-hatching.

the difficulty of determination. Fortunately, the metamorphism of the Huronian iron-formations, though widespread, has a considerable range in intensity from place to place. Gradients are relatively steep, and in the Marquette and Gogebic ranges of Michigan and the Mesabi range of Minnesota the metamorphic gradients parallel the strike so that the progressive metamorphism can be observed for definite stratigraphic units of iron-rich rock. The distribution of the iron-formations in relation to generalized areas of moderate to high-grade metamorphism is shown in Figure 1. The grade of metamorphism of the iron-formation can be established by reference to minerals such as chlorites, micas, amphiboles, and garnet, but is most readily determined by the grain size of the "chert." (The term "chert" is used in the Lake Superior

region for the silica that forms thin layers alternating with iron-rich layers in the iron-formation. The material is certainly chemical in origin—that is, nonclastic—but with a few exceptions is now entirely crystalline quartz.) The grains have a range from about 0.01 mm average diameter in low-grade metamorphic zones such as the Iron River district to 1 mm or more average diameter in the high-grade metamorphic zones such as the west end of the Marquette range. Though this criterion must be used with caution—the associated materials have a strong modifying influence—it does provide an exceedingly delicate indicator of metamorphic grade. The features of iron-formation believed to be of primary origin are described in terms of rocks from the areas of lowest metamorphic grade, as determined by the criteria of metamorphic index minerals and chert grain size.

ACKNOWLEDGMENTS

The field studies on which this paper is based have been part of a regional study by the U. S. Geological Survey of the Lake Superior iron ranges. The writer is deeply indebted to many of his colleagues for field data and critical discussion, and for laboratory assistance such as chemical analyses, X-ray identification of minerals, and preparation of photomicrographs.

THE TERM "IRON-FORMATION"

The term "iron-formation" as a rock name is one that gradually has gained general acceptance as a field term among geologists and engineers. In the literature the term has been used without definition or has been avoided, and the more cumbersome "iron-bearing formation" substituted. In the Lake Superior region the term is used in reference to a banded iron-rich rock of sedimentary origin. The rock commonly contains thin layers or laminae of chert, which alternate with layers consisting predominantly of an iron-rich mineral; and to some geologists the presence of chert is considered a necessary feature of the definition. However, in many areas—indeed, in parts of the iron-bearing strata in all areas—iron-rich rock is present that does not contain the characteristic chert layers. Commonly such rock is called "slaty iron-formation," "carbonate slate," or "ferruginous slate" and included with the iron-formation proper. In the Cuyuna district of Minnesota a considerable part of the rock that makes up the iron-formation is lacking in chert. The iron contained in iron-formation may be in the form of magnetite, martite, specularite, soft red hematite, "limonite," siderite, chlorite, greenalite, minnesotaite, stilpnomelane, grunerite, fayalite, or pyrite. It appears clear, therefore, that neither the presence of chert nor the mineralogy of the iron can enter into the definition, although both may be incorporated as modifying adjectives.

The basic feature remaining to the definition of iron-formation is the high content of primary (sedimentary) iron. A second feature, to the writer's knowledge always present, is thin-bedded or laminated structure. In this paper, therefore, iron-formation is defined as follows: *a chemical sediment, typically thin-bedded or laminated, containing 15 percent or more iron of sedi-*

mentary origin, commonly but not necessarily containing layers of chert. This is admittedly a very general definition, yet a more restricted definition would exclude some rocks now referred to as iron-formation. It encompasses the "itabirite" of South America, the "quartz-banded ores" of Sweden, the "banded hematite quartzites" of India, and the "banded ironstones" of South Africa, all of which typically consist of interlayered chert (crystalline quartz) and iron oxides. The definition does not provide a sharp distinction with the noncherty iron-rich strata of younger age; indeed, insofar as the mineralogy and content of iron is concerned, there is no real difference. In the present paper, the term is extended to include rock that normally would not be called iron-formation, such as graphitic pyritic slate, because of facies relationships to more typical varieties of iron-formation.

CHEMICAL ENVIRONMENTS OF SEDIMENTATION

Precipitation of Iron

Of the common metallic elements that precipitate to form the chemical sediments, iron is virtually alone in its ability to exist in several valence states. It may be precipitated in the ferric state as an oxide ("limonite" or hematite) or as a silicate (glauconite), in the ferrous state as carbonate (siderite) or silicate (chamosite), or as a sulfide (pyrite). As chemical precipitates—or products of diagenesis—these minerals are equilibrium products that record certain elements of the chemistry of the environment of origin. In most sediments, iron is present in such small quantity that the mineralogical differences are overshadowed by other sedimentary features. But in iron-formation, the iron is at once an indicator of environment and a major constituent. A facies classification that is based on primary differences in mineralogy, therefore, is also to a considerable extent a classification of environments.

In a recent paper, Krumbein and Garrels (57) discuss at length the major factors of importance to the accumulation of chemical sediments. The deposional environments of the iron-rich rocks clearly belong to their class of "restricted" environments, in which "physiographic, tectonic, or biologic features impose controls on circulation, oxygenation, or concentration of dissolved salts." In this type of environment, the oxidation-reduction potential, Eh, may range from positive (or oxidizing) at the surface to strongly negative (or reducing) at depth; the hydrogen-ion concentration, pH, may show a range that indicates mild alkalinity at the surface, changing to neutral or even slightly acid at depth. Concentration of dissolved salts may show wide variation, particularly when free circulation of water to the open ocean is inhibited. In contrast, in the normal marine environment, which is one of free circulation, variations in concentration of dissolved salts, acidity, and oxidation-reduction potential are relatively slight. According to Krumbein and Garrels (57, p. 3) the pH of the normal marine environment ranges from 8.4 at the surface to 7.5 at the bottom; Eh ranges from + 0.4 at the surface to + 0.1 at the bottom. In other words, in a typical marine environment of open circulation, the water is mildly alkaline and oxidizing throughout, whereas in a

restricted environment, the surface waters are alkaline and oxidizing, but the bottom waters may be reducing and acid in reaction.

Krumbein and Garrels (57, p. 9–13) also derive the approximate stability fields of pyrite-siderite-hematite, in terms of Eh and pH, by consideration of the activities of Fe‴, Fe″, $(CO_3)^{--}$, $(OH)^-$, and S^{--}, and the activity products of the compounds. The diagram showing stability fields is reproduced in

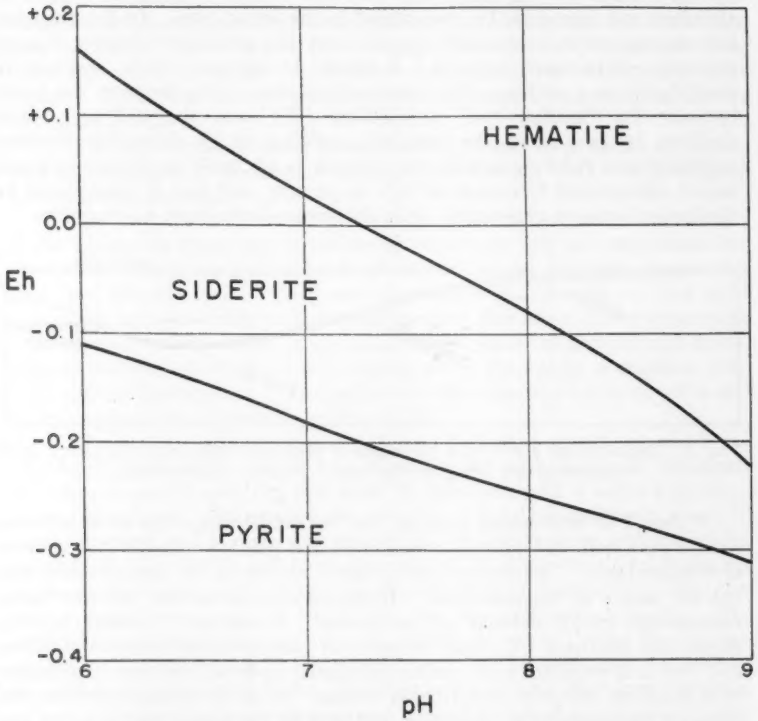

FIG. 2. Stability fields of hematite, siderite, and pyrite (from Krumbein and Garrels, 57).

Figure 2. Uncertainties are present in that the values for the activity product of "pyrite" actually are those of FeS, and those for "hematite" are $Fe(OH)_3$. Possibly of more consequence is the fact that the values used for $(CO_2)^{--}$ and $(S)^{--}$ are those of normal marine water. Nevertheless, the gross pattern of distribution in the Eh-pH field cannot be changed greatly, and the diagram serves to show the general relations between mineral phases and chemical en-

vironment. The diagram shows clearly that the form in which iron is precipitated is most strongly dependent upon the oxidation-reduction potential, Eh.

The relationships between physical environment and type of precipitate can be diagrammatically illustrated, as in Figure 3. Although it is doubtful if the pattern of precipitation indicated is ever actually obtained in nature because of complicated relationships between depth of basin, height of barrier, and details of circulation, the facies relationships between the various mineralogies are valid. On the shallow, wave- and current-swept shelf, oxygen is abundant and iron must be precipitated in the ferric state. In the stagnant bottom areas, on the other hand, oxygen is not even adequate to remove organic material, and in consequence H_2S is formed by bacterial action, and iron is precipitated as a sulfide. The intermediate zone, lying between the toxic bottoms—the "sulphuretum" of Galliher (*27*)—and the well-oxygenated shallows, is characterized by reducing conditions or by alternation between oxidizing and reducing conditions; oxygen is adequate to remove organic matter and prevent formation of H_2S in quantity, and iron is precipitated in the ferrous state, as a carbonate, or in the ferrous-ferric state, as magnetite.

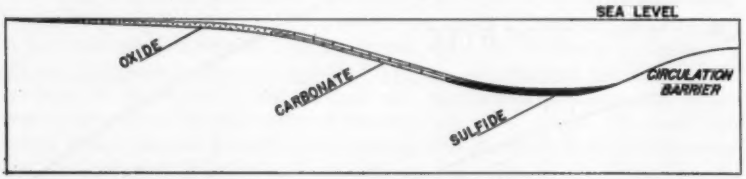

FIG. 3. Depositional zones in a hypothetical restricted deep basin in which iron compounds are being precipitated. Highly diagrammatic.

It should be emphasized, perhaps, that the relationships that exist between sulfide, carbonate, and oxide, in terms of Eh and pH, are virtually independent of concentration. The amount precipitated depends on the concentration but not the nature of the precipitate. It should also be pointed out that these relationships are for material in ionic solution. A number of workers, notably Moore and Maynard (*67*) and Gruner (*33*) have produced evidence to show that iron is transported most readily as a ferric hydrosol and that it probably is in this form that most iron reaches the sea. In all likelihood, however, the chemical environment of deposition still will be the factor that governs the type of iron mineral deposited. Consider, for example, the situation if a major stream carrying 5 ppm Fe_2O_3 emptied its load into a basin of the type illustrated in Figure 3. If it is assumed that the flow is about 120,000 cu m/sec and that the load is distributed uniformly over a semicircular area of 100 miles radius, then the increment over the entire area would only amount to about 0.1 mm Fe_2O_3 per year. (This, incidentally, is the order of magnitude of the fine laminations of the iron-formation.) If this colloidal ferric oxide settles into a reducing environment in which carbonate is the stable product, certainly it would either be taken into solution and reprecipitated as carbonate, or it

would simply be replaced by carbonate during the settling period or on the sea bottom. Only in the oxidizing environments would the colloidal oxide be permitted to accumulate without interference, and here only because it would be chemically in harmony with the ionic precipitates.

Marine vs. Nonmarine Environment

In the preceding discussion, the tacit assumption has been made that iron-formation originated in marine environments. In view of the fact that Tyler and Twenhofel (93) conclude that the Lake Superior iron-formations were deposited under essentially fresh-water conditions, this assumption bears scrutiny. The chief considerations that seem to suggest a marine origin to the writer are as follows:

1. The iron-formation and associated strata possess none of the attributes, such as predominance of coarse clastics and poor sorting, traditionally assigned to continental deposits.

2. The major iron-formation units, such as the Negaunee iron-formation of the Marquette range, are of middle Huronian age and are sandwiched between lower Huronian quartzite and dolomite and upper Huronian graywacke, slate, and volcanic rocks. The lower Huronian sequence with its thick algal dolomite is unquestionably marine, and the upper Huronian suite of graywacke and volcanics is one of the classic assemblages found in geosynclinal tracts, such as the Mesozoic of the Pacific Coast, where the weight of evidence indicates marine environment. The major iron-formations are thus overlain and underlain by strata of probable marine origin.

3. The pyritic slates associated with iron-formation proper in some areas may be of significance. In the Iron River-Crystal Falls district of Michigan, the footwall pyritic slate is a bed about 50 feet thick with a sulfur content of nearly 20 percent. This rock is preserved in a tightly folded basin about 20 miles in length and 8 miles average width. A conservative estimate of the amount of sulfur in this unit is on the order of 25 billion tons, and, inasmuch as the original extent of the bed probably was at least several times that preserved, possibly in excess of 100 billion tons was originally deposited. The availability and fixation of such a vast quantity of sulfur during the time required for deposition of a 50-foot, partly clastic bed suggests that the basin of deposition was connected to the open sea; in the absence of volcanic activity, which is ruled out by field evidence, it is difficult to conceive of another source adequate to furnish such an amount.

4. Younger iron-rich strata, such as the Clinton formation, are similar in some important respects to the Precambrian iron-formations, at least with regard to iron content and mineralogy. The marine origin of these beds is demonstrated beyond question by abundant fossils.

The major argument in favor of a nonmarine origin—or perhaps more accurately, against a marine origin—is the virtual absence of the typical calcium carbonate precipitates and the abundance of ferrous carbonate. (As will be discussed later in this paper, the actual carbonate present in iron-formation is a complex molecular mixture that commonly contains about 5 percent

$CaCO_3$ and 12 percent $MgCO_3$.) Normal sea water is essentially saturated with both Ca^{++} and Fe^{++}, and precipitation of both as carbonates probably is accomplished normally by increase in pH. But local conditions in marginal basins, even those connected to the open ocean, may deviate considerably both with respect to ionic concentrations and pH. One may reasonably postulate, for example, a marginal basin where debouching streams are of low pH, low in Ca^{++}, and high in Fe^{++}. The pH of sea water present in such a basin will be lowered so that even if the absolute Ca^{++} content remains unchanged, the water might become undersaturated with respect to that constituent. At the same time an absolute increase in Fe^{++} will take place with possible subsequent precipitation at the saturation point. Alternately, it might be assumed that the iron was transported and reached the sea in the form of colloidal ferric hydroxide, which would be precipitated by salt water. The stability of iron precipitates with respect to redox potential (Eh) would now play an important role. The solubility of iron in the surface waters of the basin is almost negligible. Using values obtained from Krumbein and Garrels (57), we may assume a pH of about 8.0 and an Eh of about $+0.2$, at which the activity of Fe^{++} plus Fe^{+++} ions is about $10^{-10.5}$, and ferric hydroxide is a stable precipitate. But as this flocculent material slowly settles, it may, in a "restricted" basin, sink into an environment more acid and less oxygenated. In such an environment a typical value for pH would be 7.5 with an Eh of -0.1, at which the activity of Fe^{++} plus Fe^{+++} is about $10^{-4.7}$, and the stable iron mineral is $FeCO_3$. This tremendous increase in solubility would cause the flocculent hydroxide to either go into solution and thus greatly increase the Fe^{++} content of the water, or, if the solubility product of $FeCO_3$ were exceeded, to be reacted with to form a ferrous precipitate. If this condition existed coincident with a slight undersaturation of Ca^{++}, the observed mineralogy of carbonate iron-formation might be accounted for.

Regardless of theoretical considerations, beds consisting dominantly of ferrous carbonate are found interbedded with normal marine limestone in some of the younger iron-rich strata of the world (see Hallimond, 40), so whatever the process, deposition of ferrous carbonate *is* possible in marine environments. This fact further suggests that the cause does not lie in profound differences between Precambrian sea water and that of more recent age. Study of minor-element distribution, such as that of Landergren (60) and of Palmqvist (69) on the Swedish iron ores, may provide more positive evidence on environmental variations.

The Problem of the Atmosphere

A number of workers have suggested that the atmosphere prevailing during the time of deposition of the Precambrian iron-formations was deficient in oxygen and rich in carbon dioxide with attendant profound modification of environmental conditions. Some have based this view on theoretical considerations regarding evolution of the earth's atmosphere. Others have seen it as a necessity to account for the ultimate origin of the vast amount of CO_2 now tied up in the limestones and dolomites; still others believe it is indicated by the fact that certain plants appear to have an inherited tolerance for CO_2

pressures several times that existing at present. The problem is an exceedingly broad one and extends into fields outside of the writer's competence. The present discussion will be confined to the evidence contained within the rocks themselves. For a more complete review of the problem, the reader is referred to a recent paper by Rubey (77), who finds that the evidence does not favor radical change in the atmosphere in recorded geologic time. Rubey suggests that much of the great volume of CO_2 necessary to account for that tied up in the limestones and dolomites is derived from volcanic emanations.

MacGregor (63) analyzed the problem in terms of the lithologic characteristics of Precambrian rocks, with specific reference to the banded iron-formations and concludes that the evidence indicates an oxygen-deficient, CO_2-rich atmosphere. His major arguments can be discussed under three headings, as follows:

1. *Widespread Carbonation of Basic Volcanic Rocks.*—This condition is undoubtedly true of many Precambrian terranes, but it is also true—though possibly to a somewhat lesser extent—of some of younger age. The formation of carbonate in submarine volcanics, according to Eskola, Vuoristo, and Rankama (26), can be attributed in part to reactions during albitization contemporaneous with extrusion; in part it can be attributed to metamorphism under conditions that give rise to the greenschist facies as shown by Eskola (25) and Turner (88). Whether or not the available evidence is statistically in favor of more intense carbonation during the Precambrian must still be considered a moot point.

2. *Predominance of Ferrous Minerals in Clastic Sedimentary Rocks.*—MacGregor presents four analyses of "calcareous graywacke" and "arkose" from Rhodesian localities (from the analyses it is evident that the "arkose" is closer to true graywacke in composition). The analyses, which show an average Fe_2O_3 content of 0.5 percent and an average FeO content of 6.1 percent, are compared with average analyses of shale and quartzite, which show a predominance of ferric over ferrous iron. But the comparison is not valid. The true graywackes, of whatever age, are essentially ferrous. The Rhodesian rocks might be better compared with, for example, the graywackes of the Mesozoic Franciscan group of California. Analyses listed for these rocks by Pettijohn (72) show them to have an Fe_2O_3 content of 0.74 percent and FeO content of 3.10 percent. However, despite the doubt concerning the particular comparisons made, there may be slight but real differences in the Precambrian rocks. Nanz (68) has shown that the available evidence suggests some progressive change in chemical composition with age; his data show that the older rocks do possess a higher proportion of ferrous iron, though in part this may be accounted for by greater degree of metamorphism.

3. *The Existence and Character of the Banded Iron-Formations.*—In considering the composition of the iron-formation, MacGregor makes the implicit assumption that the iron in these rocks is predominantly, if not exclusively, ferrous. Part of the purpose of the present paper is to show that one of the important facies of iron-formation in the Lake Superior region is one in which the original iron was in the form of ferric oxide. Later workers in South

Africa, such as du Preez (21), have developed the same view for some of the principal iron-formations of that area.

However, iron oxide could be deposited under appreciably greater CO_2 pressures than at present, though probably not under a dominantly CO_2 atmosphere except as an unstable phase. A tenfold increase in partial pressure of CO_2 (that is, $P_{CO_2} = 0.0030$ as compared with the present $P_{CO_2} = 0.0003$) would not seriously modify the oxygen pressure and the consequent Eh value of sea water, though it would cause a strong change in pH (from a present pH of 8.17 to approximately 7.4, according to Rubey's diagrams). Such a change in the atmosphere would indeed have profound effects on weathering, on transportation of iron, and on precipitation of calcium carbonate. But as seen from Figure 2, the iron precipitation is more strongly dependent upon Eh than pH; the effect of lowered pH can be overcome by an increase in the redox potential, which in terms of sedimentation means that the depositional boundary between carbonate and oxide precipitates might be moved into slightly shallower (or at least better aerated) waters. The absence of oolitic siderite in the carbonate facies of iron-formation in contrast to the prevalence of this structure in the oxide facies indicates that this boundary never reached the zone of wave action.

It seems reasonable to conclude that at the time of accumulation of the Precambrian iron-formations, the atmosphere was not greatly different from that prevailing today. Oxygen was sufficiently abundant to permit deposition of ferric hydroxide in appropriate environments in proportions at least comparable to those of iron-rich rocks of younger age. The atmosphere may well have been of quite different composition at some early stage of the earth's history, but that stage, if it existed, appears to predate the later Precambrian.

THE FACIES OF IRON-FORMATION

General

As might be expected, the lithologies encountered in the field are more complex than those suggested by generalized theoretical considerations. Four clearly defined end members—sulfide, carbonate, silicate, and oxide—make up a large part of the iron-formations. But many of the rocks were deposited in fluctuating environments, with resultant interlayering and modification of contrasting materials; others were strongly modified by diagenesis under chemical environments much different from those existing during deposition; still others owe some of their properties to clastic additions during chemical sedimentation. These factors make pigeon-hole classification of the present rock units a difficult task, but at the same time the preserved record furnishes valuable evidence with respect to relative stabilities of the minerals and the nature of the depositional and burial environments.

The sulfide, carbonate, and oxide (hematite) facies stand in well-defined theoretical and field relationships to one another. The analysis by Krumbein and Garrels (57) shows that there is little or no overlap of stability fields. This is borne out by field evidence; except for diagenetic modifications—to

be discussed later in this report—pyrite, carbonate, and hematite are mutually exclusive within a given layer of iron-formation. Also, since the chemistry of depositional environments does not change radically within short periods of time, interbedding, if present, is between adjacent types—that is, between sulfide and carbonate or between carbonate and oxide—rarely if ever between sulfide and oxide.

The silicate facies does not appear to occupy a fixed position in terms of Eh, the major environmental control. Iron silicate apparently of primary origin is present as interstitial material or as discrete layers in rocks composed dominantly of sulfide, carbonate, or oxide. However, it occurs most abundantly in rocks with layers that contain either oxide or carbonate. This association suggests that the optimum environment for its deposition is one that lies, with respect to Eh, near the boundary zone between oxidizing and reducing. But other factors, such as addition of clastic material, probably are involved in the formation of the silicate rocks. Several fairly distinct types can be recognized, but the susceptibility of primary silicate rocks to metamorphism and the development of secondary silicate rocks through metamorphic reactions in carbonate iron-formation are serious obstacles to reconstruction of original characteristics and environments of deposition.

Iron-formation that contains iron chiefly in the form of magnetite is a major rock type in the Lake Superior region. Much of this rock is believed by the writer to be of primary origin and is classed as a variety of the oxide facies. This facies—or subfacies—is not predicted by the Eh-pH diagram of Krumbein and Garrels (57), in which the only oxide indicated is hematite. Although almost certainly some magnetite in the iron-formation is formed by metamorphic reactions such as breakdown of iron carbonate or primary iron silicate, this explanation seems quite inadequate to account for some of the magnetite-banded rocks. The principal lines of reasoning that appear to support a concept of primary origin for these rocks are as follows:

1. The amount of magnetite (plus martite formed by surficial oxidation of magnetite) in most types of iron-formation bears relatively little systematic relation to the degree of metamorphism. This is noted in descriptions of the Gunflint range in Minnesota by Broderick (6, p. 440), who states: "For instance, the magnetite-rich rock of the Lower Slaty horizon were sampled in Sec. 28, where the (Duluth) gabbro is close by, and in Sec. 22, where it is further away. . . . The two samples had the same percentage of magnetite within the limits of sampling error." Concerning the Biwabik iron-formation of the Mesabi, Broderick writes: "Thus, drill cores of the unoxidized Biwabik formation, far from any igneous intrusives, show that the chief iron-bearing mineral is magnetite."

2. Layers composed almost entirely of magnetite are found interbedded with layers of iron silicate or of iron carbonate. Intimate association with these two minerals in layers less than a millimeter thick would seem to rule out the possibility of derivation by metamorphism from either constituent. Derivation by metamorphism of hematite likewise seems improbable in view of the fact that in areas of intense metamorphism such as the Republic district

of Michigan, hematite remains a major constituent—much of the banded specularite-jasper rock there contains virtually no magnetite.

3. Textural relations to be described later show that at least some of the magnetite is diagenetic in origin. Doubtless the diagenetic environments differ considerably from the depositional environments in terms of Eh, pH, and composition of the liquid phase, yet the occurrence of magnetite as a diagenetic mineral indicates that its field of stability is within reach of ordinary temperatures and pressures.

4. The magnetite-banded rocks are typically associated with silicate iron-formation, the principal varieties of which, as stated previously, appear to have been formed in an Eh environment intermediate between the oxide—that is, hematite—and carbonate zones. Such an environment might also be favorable for formation of a ferrous-ferric oxide mineral under certain conditions.

The principal features of the sedimentary facies of iron-formation are summarized in Table 1 and are described in greater detail in the pages that follow.

SULFIDE FACIES

Occurrence

The sulfide facies of iron-formation in the Lake Superior region, so far as known at present, is represented only by pyritic carbonaceous slate. Thick masses of pyrite-siderite in the Michipicoten district of Ontario are associated with banded iron-formation, but according to Hawley (45), Moore (65), and Moore and Armstrong (66), these deposits are of hydrothermal origin.

Pyritic slate is known to occur as a relatively minor rock type in several of the ranges. In the Gogebic, Marquette, and Menominee ranges of Michigan, graphitic pyritic slate occurs in the upper Huronian, stratigraphically well above the major iron-formation of the middle Huronian, and is typically associated with thin beds of carbonate iron-formation (47, 94, 4, 75, 76). In the Cuyuna district of Minnesota, pyritic rock occurs as thin layers in parts of the productive iron-formation. Sulfide-bearing graphitic slate is known to be locally abundant in outlying parts of that district, but Thiel (87) concludes that in the latter areas the sulfides (mainly pyrite and pyrrhotite) are hydrothermal.

The major occurrence of pyritic slate in the Lake Superior region probably is in the Iron River-Crystal Falls district of Michigan. Throughout this district the productive iron-formation is underlain by a pyritic black slate with an average thickness of about 50 feet (52, 53, 50, 51, 22, 73, 74).

Description

The pyritic slate, which is a sulfide facies of iron-formation, is typically a thin-banded black rock with enough free carbon to yield a distinctly graphitic streak. A typical specimen is shown in Figure 7.[1] The pyrite, even where it makes up as much as 40 percent of the rock, is for the most part not visible on unpolished surfaces except where the rock is metamorphosed. Chert layers

[1] Figures 7–33 are at back of paper.

TABLE 1

PRINCIPAL FEATURES OF THE IRON-FORMATION FACIES

	Sulfide	Carbonate	Silicate		Oxide	
			Nongranular	Granular	Magnetite-banded	Hematite-banded
Lithology	Laminated to thin-banded black pyritic carbonaceous slate. Chert rare	Thin-bedded to laminated rock consisting of alternating layers of gray chert and carbonate	Laminated light-green to greenish-black rock. Chert rare	Massive dark-green rock with irregular bedding marked by layers of chert and magnetite	Dark, thin-bedded to irregularly bedded rock consisting of layers of magnetite alternating with layers of dark chert, green silicates, or silicates plus carbonate	Thin-bedded to irregularly bedded rock consisting of alternate layers of crystalline hematite and gray chert or reddish jasper
Principal iron mineral	Pyrite	Iron-rich carbonate	Iron silicate (minnesotaite, stilpnomelane, chlorite)	Iron silicate (greenalite, minnesotaite, stilpnomelane)	Magnetite	Crystalline hematite
Subsidiary iron minerals (rarer minerals in parenthesis)	Carbonate (Greenalite)	Pyrite Stilpnomelane Minnesotaite Magnetite (Hematite)	Carbonate Magnetite	Magnetite Carbonate Hematite	Greenalite Minnesotaite Stilpnomelane Carbonate Hematite (Pyrite)	Magnetite (Carbonate)
Typical range in content of metallic iron, in percent	15–25	20–35	20–30	20–30	25–35	30–40
Distinctive features	"Graphitic"	Stylolites common	Laminated structure	Granule-bearing	Strongly magnetic	Commonly oolitic
Environment of origin	Strongly reducing, anaerobic	Reducing	Variable, but typically mildly reducing	Mildly oxidizing to mildly reducing	Mildly oxidizing to mildly reducing	Strongly oxidizing

are rare. Bedding-plane fissility is generally present, but the rock rarely possesses true slaty cleavage. The rock is not resistant to weathering and rarely crops out at the surface.

As can be seen in the illustrated specimen, the pyrite tends to be highly concentrated in some layers. The pyrite content of the layers ranges from a few percent in some to about 75 percent in others. In thin section the rocks are almost completely opaque because of the graphite and fine-grained pyrite. In polished section the pyrite is seen to form fairly well defined cubes and octahedra, as shown in Figure 8. In areas of higher metamorphic grade the pyrite grains are much larger, and many become megascopically visible. If the sulfide in the sulfide-bearing graphitic slate of the Cuyuna range is syngenetic rather than hydrothermal as postulated by Thiel (87), then under

TABLE 2

CHEMICAL ANALYSIS OF SULFIDE IRON-FORMATION

SiO_2		36.67
Al_2O_3		6.90
Fe_2O_3		—
FeO		2.35
FeS_2		38.70
MnO		0.002
CaO		0.13
MgO		0.65
Na_2O		0.26
K_2O		1.81
TiO_2		0.39
V_2O_5		0.15
P_2O_5		0.20
H_2O-		0.55
H_2O+		1.25
SO_3		2.60
Organic matter + C		7.60
(Total C	7.28)	—
Total		100.21

Locality: 10th level, Buck mine, Iron River district, Michigan (James, 50, p. 255).

certain metamorphic conditions the pyrite may be reduced to pyrrhotite, with loss of sulfur.

As indicated in Figure 3, pyritic iron-formation may grade laterally into carbonate iron-formation. Slight changes in bottom configuration or in nature of currents may cause a profound change in character of sedimentation. With slightly more oxygen available, the organic material would be removed and the iron would be precipitated as carbonate. Figure 9 shows a specimen of material in which carbonate and pyrite layers alternate. The rock clearly was deposited under boundary conditions in which the oxidation potential of the environment changed periodically, perhaps seasonally.

Composition

A complete chemical analysis of the pyritic slate is shown in Table 2. The analysis shows a pyrite content of 38.7 percent and a carbon content of 7.28

percent. (X-ray study by F. A. Hildebrand in the U. S. Geological Survey laboratory has shown that the free carbon occurs as "poorly crystalline carbon" in specimens from the Iron River district, rather than as true graphite.) A small amount of organic material is present. Partial analyses of 13 grab samples of pyritic slate from the Iron River-Crystal Falls district indicate an average pyrite content of 37.6 percent. X-ray study has revealed that greenalite is present in minor quantity in the analyzed material, but the mineral has not been observed in thin section.

Distinctive Features.

The pyritic slate is readily distinguished from all other varieties of iron-formation by its high content of free carbon. Where the rock is oxidized, as it is locally adjacent to ore bodies, the rock is dark red, but the streak is a greasy chocolate color.

Not all rocks in the region that are rich in free carbon are to be classed as sulfide iron-formation, however. Many of the graphitic slates contain only a few percent pyrite. As stated in a previous paper (50, p. 265), the high content of pyrite ". . . is the result of coincidence of two conditions: (1) abundance of H_2S on the sea bottom, and (2) exceptionally high iron content of the sea water." Either of these conditions may exist without the other, in which case pyrite will not be formed.

CARBONATE FACIES

Occurrence

Carbonate iron-formation generally has been considered the major primary lithology among the iron-rich beds of the Lake Superior region. Van Hise and Leith (95), in regard to the Deerwood iron-formation of the Cuyuna range, state (p. 223): "The member was originally cherty iron carbonate interbedded with slate;" regarding the Gogebic range (p. 231): "The cherty iron carbonate was the original rock of the iron formation." No specific statement is made concerning the Negaunee iron-formation of the Marquette range, but in the comparison of the Marquette ores with those of the Gogebic, they state (p. 276): "They are derived by the same processes, under similar conditions from cherty iron carbonate rocks which are practically identical with those of that (the Gogebic) district." The view that carbonate iron-formation is the major primary rock type in most of the ranges has been generally adhered to by most geologists.

Actually, although almost all varieties of iron-formation in the Lake Superior region contain some iron carbonate, in many of the principal stratigraphic units carbonate probably is subordinate to magnetite, iron silicate, or hematite. Clean carbonate iron-formation, consisting only of interbedded chert and iron carbonate, forms part of the "slaty" members of the Ironwood formation of the Gogebic range and part of the lower half of the Negaunee formation of the Marquette range. It is a predominant part of the iron-formation only in the Iron River-Crystal Falls district of Michigan. The

rock commonly referred to as "carbonate iron-formation" in most parts of the region is typically a mixture of iron carbonate and iron silicate in varying proportions. Tyler (91) points out that much of the iron-formation in the Marquette and Gogebic districts that has been classed as cherty iron carbonate consists chiefly of iron silicates.

Description

Carbonate iron-formation in its purer form consists only of chert and carbonate. The carbonate is fine grained, light to dark gray; the chert is dense, and, though it commonly appears nearly black on fresh break, owing to internal reflections, actually it is almost colorless—on an abraded surface, such as that of a drill core, it is nearly white. The lighter colored carbonate is found in strata that, from other evidence, appear to have been deposited under conditions approaching the oxide environment, whereas the darker colored carbonate is more commonly associated with graphitic and pyritic material; the color is probably due in part to disseminated graphite. The chert-carbonate rock is generally thin banded, with layers ranging from less than 0.1 inch to 4 inches; the average thickness of layers is roughly about 0.4 inch. Typically the layers, especially those of carbonate, show color laminations on the order of 0.01 inch in thickness. Stylolite seams, generally parallel to bedding, are common, and slump structures of contemporaneous origin are not rare. The rock is massive and does not readily part on bedding planes. Speciments of carbonate iron-formation are shown in Figures 10–11.

As a chemically intermediate type, carbonate iron-formation may grade into or be interbedded with each of the other facies of iron-formation. Interbedding of carbonate-rich and pyrite-rich layers, as shown in Figure 9, is common throughout the Iron River-Crystal Falls district. Local interbedding with hematitic rock might be expected, but it has not been observed in many areas. However, according to Stephen Royce (personal communication), the Traders member of the Vulcan iron-formation of the Menominee range, as seen in core drilled from below the level of oxidation, consists of cherty carbonate iron-formation with layers of specularite. In the James mine of the Iron River district the shaft stations at the lower levels are in unoxidized carbonate iron-formation. At the sixth level a thin layer of blue hematite, about an inch thick, could be traced for nearly 100 feet. Figure 12 is a photograph that shows this layer interbedded with iron carbonate. The carbonate, judging from its reaction to acid, is considerably more calcitic than the normal carbonate of the iron-formation. More typically, however, the carbonate-chert rock grades into a complex silicate-magnetite-carbonate-chert rock such as that which makes up much of the iron-formation of the Mesabi, Gobebic, and Marquette ranges. As stated earlier, these magnetite-silicate rocks appear to be the products of environments lying between the carbonate and oxide (hematite) zones. Rock intermediate between the carbonate and silicate-magnetite types are illustrated in Figure 19; the finely laminated structure is typical.

Tyler (91) refers to the silicate-carbonate rocks as "metamorphosed iron

formation," and considers the stilpnomelane and minnesotaite to be formed either by metamorphism of an original silicate (greenalite) or by reaction between primary carbonate and chert. However, the absence of silicate in many layers of mixed chert and siderite adjacent to layers of practically pure silicate leads the present writer to the view that the carbonate-chert reaction is of minor importance, at least in areas of low-grade metamorphism, and that almost all the present content of low-grade silicates (greenalite, chlorite, stilpnomelane, and minnesotaite) is primary or derived from original silicate material. Within the width of a single thin section cut normal to the bedding of a rock composed of chert, carbonate, and silicate, one may observe pure

TABLE 3

CHEMICAL ANALYSES OF CARBONATE IRON-FORMATION

	A	*B*	*C*	*D*	*E*	*F*	*G*	*H*
SiO_2	24.25	32.2	42.37	26.97	46.46	39.52	31.84	30.15
Al_2O_3	1.71	1.5	n.d.	1.30	0.24	0.80	2.09	0.40
Fe_2O_3	0.71	0.6	1.09	2.31	0.64	1.05	26.3 Total	5.27
FeO	35.22	31.6	31.41	39.77	26.28	31.67	26.3 iron	31.34
MgO	3.16	2.8	2.48	1.84	3.10	2.88	3.80	4.77
CaO	1.78	1.6	0.50	0.66	1.87	0.59	1.49	1.25
Na_2O	0.04	—	n.d.	0.09	n.d.	n.d.	n.d.	—
K_2O	0.20	0.2	n.d.		n.d.	n.d.	n.d.	—
H_2O	0.21	0.2	n.d.	0.61	1.15	1.13	1.80	0.36
TiO_2	0.00	0.0	n.d.	n.d.	tr.	0.15	0.12	0.02
P_2O_5	0.91	0.8	n.d.	n.d.	0.13	0.02	0.83	0.67
CO_2	27.60	24.8	21.80	26.20	19.96	21.15	19.40	24.65
MnO	2.11	1.9	n.d.	0.29	0.21	1.00	2.35	2.27
C	1.96	1.8	n.d.	n.d.	n.d.	n.d.	n.d.	—
S	n.d.	n.d.	n.d.	n.d.	0.05	—	n.d.	n.d.
Total	99.86	100.0	99.65	100.06	100.09	99.96		101.15
Total Fe	27.87	25.0	25.16	32.40	20.87	25.34	26.3	28.06

A. Banded chert-carbonate iron-formation, Iron River district, Michigan (James, 50).
B. Analyses A recalculated to total iron content of 25 percent (James, 50).
C, D. Carbonate iron-formation, Marquette district (Van Hise and Bayley, 94).
E. Carbonate iron-formation, Gunflint district, Minnesota (Irving and Van Hise, 49).
F. Carbonate iron-formation, Gogebic range. Average of analyses II, III, IV, and V given in Irving and Van Hise (49).
G. Carbonate-silicate iron-formation, Crystal Falls district, Michigan (Pettijohn, 74).
H. "Fine grained, light gray, siliceous iron carbonate rock from Kennedy Mine, near Cuyuna, Minn." (Harder and Johnston, 42, p. 120).

layers of each mineral, layers of mixed chert and carbonate, mixed carbonate and silicate, or mixed chert and silicate. Further appraisal of this problem is made later in this paper, under the discussion of the silicate facies.

Composition

Eight analyses of carbonate iron-formation are given in Table 3. Inasmuch as iron-formation is a banded rock, with layers of contrasting composition, a truly representative sample is difficult to obtain. Analysis A, for example, contains too much iron and so has been reduced to a total iron content

of 25 percent, which is the average indicated by thousands of partial analyses of the rock in that district. Analysis D with a total iron content of 32.40 percent is probably also too high in iron to be representative, whereas E, with a content of 20.87 percent, is too low.

In Table 4 are listed the calculated mineral compositions of the analyzed materials. The calculated composition is designed to conform in general with observed modes, but several assumptions are made. Most of these assumptions are not entirely valid but are necessary because of the fact that elements such as Mg and Fe are shared by more than one mineral. Except for analysis G, explained separately below, it is assumed that Ca, Mg, and Mn are combined as carbonate. The remaining CO_2 is assigned to iron carbonate. The Fe_2O_3 is disregarded if less than 1 percent; if more it is calculated as magnetite, which is an observed mineral in some of the unoxidized rocks. The excess FeO, after satisfying the requirements of the CO_2 and Fe_2O_3 content, is calcu-

TABLE 4

MINERALOGICAL COMPOSITIONS, IN WEIGHT PERCENT, AS CALCULATED FROM ANALYSES IN TABLE 3

	A	*B*	*C*	*D*	*E*	*F*	*G*	*H*
Carbonate	70.0	62.5	55.3	67.5	49.6	53.4	48.3	60.0
Chert	24.2	32.0	42.4	24.9	43.8	32.8	20.1	26.8
Silicate	—	—	—	1.0[1]	5.4[1]	11.0[1]	22.5[2]	4.6[1]
Magnetite	—	—	1.6	3.1	—	1.5	—	7.6
Kaolin[3]	—	—	—	2.3	—	2.0	1.7	1.0
Graphite	2.0	1.8	—	—	—	—	—[4]	—
Miscellaneous[5]	3.8	3.7	0.7	1.2	1.2	—	7.4	—

[1] Calculated as minnesotaite.
[2] Calculated as stilpnomelane.
[3] Not known to be actually present in the rocks.
[4] Known to be present, but not determined in the analysis.
[5] Includes pyrite, phosphate, hematite, micas.

lated as minnesotaite containing 52 percent SiO_2 and 35 percent FeO. The Al_2O_3 is assumed to occur as kaolin. All other components, such as phosphate, pyrite, hematite, and micas, are lumped under miscellaneous.

The assumption that all the silicate in D, E, and F is minnesotaite may not be entirely correct, as stilpnomelane and greenalite are common associated minerals. The composition assumed for the minnesotaite is only approximate, and the presence of several percent MgO and H_2O in its composition is ignored. However, neither of these items introduces significant error because of the relatively small amounts involved. The Al_2O_3 may well be present as phosphates or as silicates, as no kaolin is positively identified in the unaltered rocks.

The calculation of the mineralogical content for analysis G is especially difficult because FeO was not separately determined on account of the presence of organic material. However, "soluble" silica was determined to be 10.11 percent, and, because the silicate is known to be stilpnomelane with average gamma index of about 1.665, it is possible to calculate the approximate

silicate content. The resultant composition, after the requirements of carbonate and stilpnomelane are filled, leaves a considerable excess of Fe, which is chiefly in the form of pyrite.

Some systematic variation is to be noted. The two samples containing free carbon (A and G) are high in manganese. They contain pyrite (not separately listed) but no magnetite, whereas the reverse is true for most of the nongraphitic rocks.

The approximate percentage composition of the carbonate is given in Table 5, as calculated from Table 3.

No well-defined relationships are evident; with increase in the iron component, each of the other components shows a general but rather erratic decrease. The assumption is made that the carbonate material is a single phase,

TABLE 5

APPROXIMATE COMPOSITION OF CARBONATE, IN WEIGHT PERCENT, FROM CARBONATE IRON-FORMATION

	A–B	*C*	*D*	*E*	*F*	*G*	*H*
$CaCO_3$	4.6	1.6	1.8	6.7	2.1	5.6	3.6
$MgCO_3$	9.5	9.4	5.8	13.1	11.4	12.6	16.4
$MnCO_3$	4.9	—	0.7	0.6	3.0	7.7	7.4
$FeCO_3$	81.0	89.0	91.7	79.6	83.5	74.1	72.6

which may not be correct. The data given by Kulp, Kent, and Kerr (58) indicate that complete solid solution obtains between Fe, Mn, and Mg, but that substitution in the Ca-Mg, Fe series is very limited. Whether the substitution limit is exceeded by the amount of Ca present in the rocks here described is not known at present.

Distinctive Features

Stylolite seams are common in the purer varieties of carbonate iron-formation, as shown in Figures 10–11. Typically, the stylolites are of small amplitude—⅛ to ¼ inch—with columns normal to the bedding. These structures are commonly preserved even where the rock is completely oxidized.

Recognizable clastic material is exceedingly rare, except in silicate-rich layers. The true granule or oolith [2] structure composed of discrete rounded or ellipsoidal grains does not occur in carbonate iron-formation, although a granular texture that consists of tightly packed spheres of carbonate is present in some rocks (50), and, in others, diagenetic modifications of primary ooliths or granules of oxide or silicate have yielded granule-like grains of coarse carbonate as shown in Figure 33. The absence of oolitic structures in the iron carbonate rocks at first glance appears anomalous, in view of the common presence of oolitic structure in limestones and in other varieties of iron-formation.

[2] Distinction will be made in this report between *granules* and *ooliths*. The oolith is characterized by concentrically arranged material, whereas the granule is structureless or contains nonconcentrically arranged material. A rock containing ooliths is referred to as having oolitic structure, or it may be referred to as an *oolite*.

This absence may be explained by considering the type of environment under which an oolitic rock accumulates. All recent workers on the younger ironstones have pointed out the evidence for shallow water agitated by waves and currents during deposition of the ferruginous oolites. This evidence is most clearly shown by rocks such as illustrated by Cayeux (12, 14) in which the ooliths are built around cores of waterworn shell fragments and by those in which considerable chemical heterogeneity exists (41, p. 79–82). This environment is certainly one of high oxidation potential, in which ferrous carbonate would not be a stable deposit. Precipitation of calcium carbonate, on the other hand, is unaffected by the oxidation potential, and this material may accumulate within the zone of wave action and consequent good aeration.

OXIDE FACIES

General

Oxide iron-formation has not generally been considered an important primary lithology in the Lake Superior region. As quoted earlier in this report, the classic view of Van Hise and Leith (95) is that the iron-formations consisted chiefly of cherty iron carbonate. However, at the time of the monograph studies (1885–1912), the concept of widespread primary deposition of hematitic rocks, as exemplified by some of the unmetamorphosed ironstones of Paleozoic and Mesozoic age, had not been completely developed or accepted; at that time many held that the oolitic hematites, such as the Clinton formation of New York, were formed by postconsolidation replacements of oolitic limestone (44).

The concept of primary origin of hematite for oolitic formations of younger age was expressed in successive papers by C. H. Smyth (80, 81, 82) and gradually gained general acceptance. The classic study made by Hayes (46) under Smyth's direction on the Wabana ores of Newfoundland established firmly the primary deposition of hematite, chamosite, siderite, and pyrite. Hallimond (40) and Taylor (86) show that iron oxide is a common original constituent of the British ironstones of Jurassic age, originating either by primary deposition or by sea-bottom oxidation of other iron minerals. More recently, Alling (2) returned somewhat to the views of Cayeux (12, 13, 14) and postulates that the oolitic ores of the Clinton formation (New York) were formed by a series of diagenetic or sea-bottom replacement of calcium carbonate. The hematite, however, is still to be considered a preconsolidation product.

Castaño and Garrels (11) carried out a series of experiments to test the hypothesis of replacement of calcite by iron oxide. In reference to deposition of the Clinton ores they point out that the field evidence indicates wave action and consequent high oxidation potential during deposition. They state (11, p. 768): "The iron bearing solutions would then precipitate the iron as ferric oxide, and, if the experimental work is a guide, the iron oxide would replace the calcium carbonate as well as precipitate freely in the water."

The Precambrian banded ironstones of South Africa, which consist of banded chert and finely crystalline hematite, were regarded as originally having

been sideritic cherts by Wagner (96, p. 64), who states: "As to the nature of this carbonate, we have no direct evidence. From the fact, however, that in the Lake Superior region complete gradations can be made out between banded ironstones identical in every respect with those under description and banded sideritic cherts from which the ironstones have clearly been derived, it appears reasonable to assume that the ironstones in their original form were banded sideritic cherts . . ." but in a recent, more detailed account of some of these ironstones in Transvaal, du Preez (21, p. 358) concludes that ironstone was not originally sideritic, but: "Chemical precipitation of the silica and iron hydrosols resulted in the deposition of the banded ironstone formation."

The Precambrian "quartz-banded ores" of central Sweden, which are hematitic rocks virtually identical with some types of Lake Superior iron-formation, are now regarded by Geijer (28) as primary oxide sediments.

The "banded siliceous hematite" of Morro do Urucum, Brazil, consists of banded silica and hematite. Dorr (18, p. 21) states: "All the evidence now available indicates that the rocks of the Jacadigo series are primary sediments. . . . No evidence is known to indicate that the iron was originally deposited as carbonate or silicate."

The preceding discussion makes it clear that primary hematite iron-formation is a recognized major rock type in both Precambrian and younger rocks in many parts of the world. Although a number of workers such as Grout (31), Gruner (33), and Gill, (29, 30) have stated or implied that some of the hematitic rocks of the Mesabi and Gunflint ranges are of primary origin, the concept has not received general notice or acceptance.

Corroborative evidence to support the view that magnetite may be a primary mineral of major importance is not as plentiful as that available regarding hematite. Within the Lake Superior region, both Broderick (6) and Gruner (39) point out that the dominant mineral in the unoxidized and relatively unmetamorphosed Biwabik iron-formation of the Mesabi is magnetite. Edwards (23) has shown that the iron in the banded "hematite-quartzite" of the Middleback Ranges of Australia is dominantly martite—that is, hematite pseudomorphous after magnetite. Residual magnetite is present, and Edwards states (23, p. 172): "These minute cores of magnetite are so widespread throughout the hematite ore bodies as to indicate that a large proportion, if not the bulk, of the hematite was originally magnetite." Miles (64, p. 197), in describing the iron-formations in Western Australia, writes: "The evidence so far obtainable, then, indicates that the existing primary iron oxide of the banded iron formations of the older Precambrian Series in Western Australia, is magnetite." However, both Edwards and Miles point out that the magnetite in turn may have been derived from some preexisting iron mineral.

Magnetite is a relatively scarce mineral in the unmetamorphosed younger ironstones of the world, although Taylor (86, p. 20) describes the local presence of magnetite as either a minor or dominant constituent of ooliths in the flat-lying Northampton sand ironstone (Jurassic age) of England. Hallimond (40) refers to the Northampton occurrence and also to others in Great Britain. In regard to magnetite in some of the Welsh ironstones he states (p. 25):

". . . it seems clear that the magnetite is a primary constituent." Spiroff (84) has shown that magnetite may be formed at ordinary temperature and pressure from mixtures of ferric and ferrous materials. In a critical examination of the so-called "melinkowite," a reported magnetic iron sulfide in glauconitic shale of Miocene age in southeast Russia, Berz (5) reports that the rock actually contains pyrite and magnetite, rather than a single magnetic sulfide. He reviews the occurrence of magnetite in unmetamorphosed sediments, citing Gumbel, Collet and Lee, and Cayeux to show that magnetite of primary origin is not uncommon in many ironstones and is abundant in some. He notes that chamosite or glauconite may be so heavily charged with magnetite as to be opaque. Berz concludes that magnetite may be formed either directly from ferric oxide in the presence of decomposing organic material or as a ferrous oxide that is converted to magnetite. Krotov (56) describes magnetite in sedimentary pisolitic-oolitic deposits in the Urals. He considers the magnetite to have originated by diagenesis of ferric oxides in the sediment, in the presence of decaying organic material.

Brown (7) reviews the occurrence of low-temperature magnetite and concludes (p. 147): "Magnetite has formed at a number of localities by natural supergene or superficial low-temperature processes. . . . Magnetite forms in such situations only when a favorable, and rather delicate balance exists between oxidizing and reducing tendencies."

Yet the formation of magnetite does not seem to be a normal process, as pointed out by Taylor in his discussion of British ironstones (41, p. 80), who states: "The rare and locally developed ooliths with zonal development of finely divided magnetite present a special problem. The development of magnetite as a low temperature mineral is evidently exceptional and is not a normal result of partially reducing conditions." Taylor bases this view on the absence of magnetite in ooliths that show an alternation of chamosite (a ferrous mineral) and limonite, which presumably reflect an alternation between oxidizing and reducing conditions at the time of origin. The abundance of magnetite as an apparent primary mineral in the iron-formations might conceivably be interpreted to indicate *both* low O_2 and CO_2 atmospheric pressures at the time of deposition—low O_2 to inhibit formation of a higher oxide, and low CO_2 to inhibit formation of ferrous carbonate.

Hematite-Banded Rocks

As mentioned previously, primary hematite rock is a minor lithology in the Mesabi range (39) and in the Gunflint range (29). Gruner (39) describes hematite interbedded with siderite and states (p. 31): "Hematite occurring with siderite in this manner is unquestionably primary." In the Michigan ranges the jaspilite that forms the upper part of the Negaunee iron-formation in the Marquette district is considered an example of primary hematite iron-formation, as is the Curry member of the Vulcan iron-formation of the old Menominee range. Both units consist of thin-banded chert and finely crystalline hematite. In the Republic district, the jaspilite of the Negaunee iron-formation retains its composition and thin-banded character but has been

metamorphosed to a rock consisting of granular quartz and coarse specularite.

In general, the hematite-bearing rock here referred to as a primary lithology has been considered the product of premetamorphic weathering of carbonate iron-formation by most geologists. This view, outlined by Van Hise and Leith (95) and reaffirmed by Leith (62) is stated by Royce (75, p. 79) as follows: ". . . much of the oxidation of the original Middle Huronian carbonate formations to hematites took place in immediately pre-Upper Huronian time . . . mountain-building movements and intrusions which took place in Upper Huronian and later time have resulted in specularizing by anamorphic action such hematites and limonites as were formed prior to the Upper Huronian." However, the known presence of hematite layers in unoxidized carbonate rock, as shown in Figure 12 and as described by Gruner (39), demonstrates that hematite was definitely deposited in some areas. There is, therefore, no *a priori* reason for not assuming that at least part of the premetamorphic hematite is of primary origin; certainly such a facies is a predictable one in the region. Aside from these considerations, angular fragments of interlayered hematite and chert are abundant in the Goodrich quartzite, which overlies the Negaunee iron-formation of the Marquette range with slight unconformity. To have yielded such fragments, the Negaunee must have been well lithified; if it had been a weathered carbonate, the secondary oxides would have washed out, leaving just plates of chert. Furthermore, the preservation of exceedingly fine bedding and of knife-edge contacts between jasper and hematite (Fig. 13) in the upper part of the Negaunee would not seem to be characteristic of weathered rock.

Description.—The rocks assigned to the hematitic oxide facies consist principally or entirely of interbedded chert and hematite. The jaspilite of the Marquette range, a spectacular example, consists of interlayered, even-bedded reddish jasper and fine-grained steel-gray hematite; the latter, on weathered surfaces, assumes an even higher luster than on the freshly broken surface. The appearance of the jaspilite in outcrop is shown in Figure 13, although a black and white photograph does not do justice to this colorful rock. The layers are typically 0.1–0.5 inch thick, but layers up to 4 inches are not uncommon. The layers themselves are laminated, the laminae being on the order of 0.01 inch in thickness. The Curry member of the Vulcan iron-formation is thin-bedded chert or jasper and hematite similar in appearance but less brilliant in color than the jaspilite. In part, the bedding is wavy rather than even, with individual layers showing pronounced pinch-and-swell.

In many places the hematitic rock is oolitic. The ooliths, where undeformed, are spherical or ellipsoidal in shape and average 0.5 mm in diameter. They are readily visible on a wetted surface. The oolitic structure is most evident in the jasper layers, the hematite accounting for the reddish color. These ooliths commonly consist mainly of silica, but the structure is defined by the hematite (Figs. 22–27). All gradations can be seen between ooliths with only a rim of hematite to granules of solid hematite. In many areas, especially near the axial regions of folds, the ooliths are strongly deformed and now consist of drawn-out lenticles (Figs. 23–24).

The oolitic structure is not limited to the jasper layers. It is present in hematite layers as well, but the ooliths are much less distinct, being scarcely visible in hand specimens except on well-polished surfaces. Very likely oolitic structure originally was general throughout the rock but is less readily preserved in the hematite-rich bands.

Hematite rock containing dispersed silicate, magnetite, and carbonate is a common though not abundant rock type, and "mixed oolite" containing ooliths or granules of both hematite and greenalite forms some layers in the upper slaty beds of the Gunflint range. This material is illustrated in Figure 25. Hematitic iron-formation is interbedded with carbonate that is calcitic or dolomitic in composition in parts of the Curry member of the Vulcan iron-formation of the Menominee range. A typical specimen is shown in Figure

TABLE 6

COMPOSITION OF OXIDE IRON-FORMATION

A		B	
Fe	36.26	SiO_2	40.1
Phos	0.30	Al_2O_3	0.8
SiO_2	40.15	Fe_2O_3	50.1
Mn	0.13	FeO	1.6
Alumina	0.77	MgO	2.0
Lime	1.39	CaO	1.4
Magnesia	1.97	Na_2O	n.d.
S	0.009	K_2O	n.d.
Loss on ignition	2.60	P_2O_5	0.7
		MnO	0.2
		S	—
		CO_2	2.6
			99.5

A. Analysis of shipments from Millie mine, Iron Mountain, Mich., for 1936. Sample dried at 212° F.

B. Analysis A recast in terms of standard oxides. The 36.26 percent content of iron is divided arbitrarily into 35 percent ferric and 1.26 percent ferrous. All loss on ignition assumed to be CO_2.

14. The composition of the carbonate has not been determined accurately, but partial analyses indicate that it is low in iron. It effervesces readily in cold dilute HCl, so it is probably calcitic.

Composition.—The writer has not been able to locate in the literature a single complete analysis of rock that definitely can be called primary oxide iron-formation. However, the commercial analysis of "siliceous ore" shipped from the Millie pit on the Menominee range is available (Lake Superior Iron Ore Association, 59). This pit is in the Curry member of the Vulcan iron-formation at Iron Mountain, Mich. Examination of the pit indicates that the rock has not been oxidized—layers of iron-rich chlorite and of dolomitic carbonate show no alteration. The analysis, which represents 13,417 tons mined in 1936 by the North Range Mining Co., is given in column A, Table 6. The analyzed material had been dried at 212° F. In column B the analysis is converted to standard oxides. The iron is arbitrarily divided into 35 percent ferric and 1.26 percent ferrous. This is done to permit calculation of a small

percentage of magnetite, which is an observed minor constituent. As the total for the reconstructed analysis is close to 100 percent, the possible error is relatively small. "Loss on ignition" is given as CO_2, inasmuch as carbonate is the only observed mineral with volatile content (except for a very minor amount of chlorite). The magnetite is present in part as intergrowths with the hematite and in part as discrete grains in the calcitic and silicate-rich layers. A small amount, probably not more than 1 percent, of a fine-grained greenish chlorite is present. The mineral is slightly pleochroic, fibrous or platy, and resembles the greenish varieties of minnesotaite except that the birefringence is very low. Its identity as a chlorite has been confirmed by X-ray examination. The mineral content indicated by the recast analysis is approximately 47 percent hematite, 40 percent quartz (chert), 5 percent dolomite, 5 percent magnetite, and 3 percent miscellaneous (phosphate, kaolin, chlorite), by weight.

Distinctive Features.—If the premetamorphic hematite is accepted as primary rather than a product of post-iron-formation oxidation, then the presence of specularite as a major constituent is certainly a distinguishing feature of oxide iron-formation. Introduced specularite of hydrothermal or metamorphic origin is not uncommon, but it is small in quantity and generally easily distinguished on the basis of occurrence.

A distinctive property of this lithology, aside from its specularite content, is oolitic structure (Figs. 22–27). This structure was recognized as a criterion for separating the Curry member from the Traders member of the Vulcan iron-formation on the Menominee range by Bayley (4, p. 342), although a different interpretation was made of its significance. The oolitic structure of hematite has been noted by Gruner (39) in the rocks of the Mesabi range and by Gill (29) for the rocks of the Gunflint range. In some areas of hematite iron-formation, the oolitic structure is found throughout, whereas in others it is found only in some layers. Locally the ooliths may be so deformed as to be virtually nonrecognizable. The structure is preserved, however, in many areas of high-grade metamorphism, as in the Felch district, where the metamorphism was not accompanied by strong shearing.

Magnetite-Banded Rocks

The magnetite-banded rocks appear to be one of the major lithologic types in the region. They are especially characteristic of parts of the iron-formations of the Mesabi and Gogebic ranges. At or near the surface, the magnetite is altered to martite, and complete leaching or replacement of the chert and silicates gives rise to the soft blue ores produced from certain beds. The middle part of the Negaunee iron-formation contains some zones of magnetite-banded rock, and part of the Traders member of the Vulcan iron-formation of the Menominee range is probably of this facies.

In many respects, the magnetite-banded facies of iron-formation is the most difficult to separate as a specific type because of its intimate association with silicate or silicate-carbonate rocks into which it may grade imperceptibly.

Description.—The most common lithologic type of magnetite-banded iron-formation is the rock generally known as "wavy-bedded iron-formation" on the Mesabi and Gogebic ranges. Gruner, who uses the terms "irregularly banded taconite" and "wavy-banded taconite," describes the Mesabi rock as follows (39, pp. 34–35): "This taconite differs from the banded taconite, in which the banding is parallel, by the irregular arrangement of the bands which are often not continuous, but pinch out or split, and curve in the most irregular manner or end abruptly. In the iron-rich portions of the formation the bands are dark gray to black in color and consist of magnetite mixed with carbonate. Between the dark bands lighter colored lenses and irregular areas of chert or carbonate are seen. Magnetite or hematite is sprinkled through these lenses and areas, giving a kind of 'salt and pepper' effect. The bands hardly ever exceed two inches in width except on the East Mesabi, and generally are less than one inch wide." Rock virtually identical to that described makes up much of the "wavy-bedded" Plymouth and Norrie members of the Ironwood iron-formation of the Gogebic, although minnesotaite generally is present in the lighter colored layers. A typical specimen is shown in Figure 15. Similar rock is present in the middle part of the Negaunee iron-formation of the Marquette district; some of this material contains green layers in which minnesotaite is the major constituent. In nearly all rocks of this type examined, the carbonate between magnetite layers is present as scattered grains that are much coarser than other constituents of the rock; in thin section, some of these grains show rhombic outlines; but the rounded outlines of others with a single crystallographic orientation suggests diagenetic replacement of granules of some other mineral, possibly iron silicate.

The mineralogic combinations described above also form laminated or thin-bedded rocks quite different in appearance from the "wavy-bedded" iron-formation, as shown in Figure 16, which is a photograph of a specimen from the eastern part of the Gogebic range. The magnetite, in layers as much as half an inch thick, is interbedded with light-green material that in thin section is seen to be composed of mixtures of minnesotaite, chert, and carbonate. The rock illustrated in Figure 17 is laminated, with magnetite plus stilpnomelane in varying proportions making up most of the layers. A magnetite-jasper rock of striking appearance is found as a minor lithologic type in the Gogebic range. In this rock, thin layers and laminae of magnetite alternate with bright-red jasper that consists of chert with disseminated fine plates of hematite. The jasper layers fray out into the magnetite-rich parts of the rock with irregular but actual truncation of bedding. Rather rarely, the jasper layers contain oolitic hematite; such layers are more irregular in thickness than those that are nonoolitic. The hematite ooliths are commonly partly to completely replaced by coarse-grained magnetite and carbonate and by minnesotaite. The mutual relations are interpreted to mean that the primary hematite was unstable in the bottom or burial environment and was altered to more ferrous compounds by diagenesis.

Composition.—The magnetite-banded rock that makes up much of the lower cherty division of the Biwabik iron-formation of the Mesabi was sam-

pled by Gruner (39) and analyzed by W. E. Apuli in the Mines Experiment Station of the University of Minnesota. In column A, Table 7, is given the weighted average of 5 analyses, given in Gruner's Table 7, that represent 110 feet of drill core; in column B the weighted average of 6 analyses, given in Gruner's Table 8, representing 110 feet. Magnetite is by far the dominant iron-bearing constituent, making up about a third of the rock by weight. Other minerals present are iron silicates, carbonate, and chert.

Distinctive Features.—Abundance of magnetite in rocks that are essentially unmetamorphosed, as indicated by the fine grain of the chert and the presence of such low-grade minerals as greenalite and minnesotaite, is a valid criterion that should serve to separate the primary magnetite rock from magnetite-

TABLE 7

COMPOSITION OF MAGNETITE-BANDED IRON-FORMATION

	A	B
SiO_2	48.75	48.28
Al_2O_3	1.70	0.23
Fe_2O_3	30.31	22.47
FeO	13.62	17.79
MgO	1.48	2.94
CaO	0.89	1.46
Na_2O	n.d.	n.d.
K_2O	n.d.	n.d.
H_2O+	1.55	1.36
CO_2	2.77	5.05
P_2O_5	0.07	0.07
C	0.04	0.08
Total Fe	31.08	28.61

bearing rocks that are products of later metamorphism. The magnetite-banded rocks are typically, almost invariably, associated with the silicate facies, and every gradation can be found between the two.

SILICATE FACIES

General

The silicate facies of iron-formation was recognized as a primary lithology of major importance by the early students of the Mesabi range. Spurr (85) described the silicate rock and concluded that the iron mineral was a variety of glauconite. Leith (61) showed that fundamental differences exist between glauconite and the iron mineral and coined the term "greenalite." Subsequent work by Gruner (33–39) provides much detailed information on the nature and distribution of the iron silicates of the Mesabi, and the 1946 paper (39) contains an excellent summary of the properties and composition of most of the iron silicates found in the region. In many ways, the silicate rocks are the least understood of the iron-formation lithologies. Precipitation of silicate does not appear to have been as simply controlled as that of sulfide, carbonate, or oxide; and it seems likely that there is more than one mode of origin. Furthermore, the silicate rocks are exceedingly susceptible to metamorphism; except for greenalite, which is almost certainly of sedimentary origin, some

doubt exists whether the present silicates are of primary or metamorphic derivation.

Minerals other than chert, carbonate, and magnetite that are of major importance in the silicate facies are discussed below. Silicates such as grunerite, that occur only in the higher grade metamorphic zones, are not given.

Greenalite.—First recognized as a mineral species by Leith (61). Greenish, more or less isotropic in thin section; index approximately 1.67; birefringence (in mass) very low. Gruner (35) has shown that it has the structure of serpentine. No completely satisfactory analysis has yet been made because of the intimate mixture of other minerals. The analysis given by Leith shows it to be of the approximate composition:

	Percent
SiO_2	32
Al_2O_3	1
Fe_2O_3	23
FeO	29
MgO	5.3
H_2O	9.5

Jolliffe (54) obtained a composition considerably different from that given by Leith. Gruner, however, considers that Jolliffe's values are probably incorrect because of contamination by other minerals, including minnesotaite, which was established as a separate species subsequent to Jolliffe's work. Greenalite apparently is a direct precipitate, but the conditions governing its deposition are not clearly understood. Castaño and Garrels (11) suggest that the somewhat similar chamositic rocks of iron oolite of the Clinton formation are direct precipitates under mildly oxidizing conditions with accompanying low pH. The greenalite commonly is in the form of irregularly shaped granules but may occur as disseminated groundmass material.

Stilpnomelane.—First recognized in the Lake Superior region by Grout and Thiel (32) as a constituent of veins cutting the iron-formation. Established as a major constituent of iron-formation by Gruner (36, 37). A flaky mineral, generally dark, fine-grained. In thin section resembles biotite; strongly pleochroic in either brown or green; indices and pleochroism vary with $FeO:Fe_2O_3$ ratio (Hutton, 48).

The approximate range in composition is as follows:

	Percent
SiO_2	42–48
Al_2O_3	4–7
Fe_2O_3	4–31
FeO	2–20
MgO	2–7
Na_2O+K_2O	0–3
H_2O	6–8

Stilpnomelane typically contains a small but notable content of the alkalies, principally potash. It is further distinguished from greenalite and minnesotaite by its higher content of alumina.

Stilpnomelane is a widespread constituent of certain schists in New Zealand, where it has been shown by Turner and Hutton (89) and by Hutton

(48) to be a product of low-grade metamorphism. Tyler (91) attaches similar significance to its occurrence in the Lake Superior region, whereas Gruner (39) considers the mineral to be a primary constituent.

Minnesotaite.—Established as a mineral species by Gruner (38); now recognized as a major constituent in many iron-formation rocks. Light greenish gray, soft, fine-grained. In thin section, forms colorless to light-green aggregates of fine fibers and plates, non-pleochroic or only slightly so; birefringence about equal to sericite. The composition given by Gruner is as follows:

	Percent
SiO_2	51.29
Al_2O_3	0.61
Fe_2O_3	2.00
FeO	33.66
MgO	6.26
CaO	none
Na_2O	0.08
K_2O	0.03
TiO_2	0.04
H_2O+	5.54
H_2O-	0.24

Like the stilpnomelane, minnesotaite is thought by Gruner to be of primary origin.

Chlorite.—Iron-rich chlorite (aphrosiderite and possibly thuringite) has been shown by the author (50) to be a major constituent of the iron-formation rocks of the Iron River district and is also abundant in the adjacent Crystal Falls district. It occurs as exceedingly fine-grained material, typically associated with iron carbonate and magnetite. The chlorite, which appears to be virtually limited to those iron-rich rocks that contain recognizable clastic material, is considered by the writer to have originated by reactions between iron-rich sea water and the finer fractions of a clastic load.

The composition of the various primary or low-grade metamorphic iron silicates, in terms of their silica, iron, and alumina content, is indicated in Figure 4. The chlorite shown is thuringite, which is close to aphrosiderite in composition. Chamosite, the primary iron silicate of most ironstones of younger age, is clearly much richer in alumina that most of the iron silicates of the Lake Superior region; the generally low values of alumina for the silicate facies of the iron-formations indicate that the chamosite probably was not present, at least in abundance, in the original deposits.

Of the silicates present in iron-formation, only greenalite is certainly of primary origin. Regarding minnesotaite and stilpnomelane, the evidence is conflicting. The iron-formation of the Gunflint range of Ontario probably is the least metamorphosed of the iron-formations in the Lake Superior ranges, as shown by the fact that much of the silica is still chalcedonic. In these rocks, greenalite is the major silicate, stilpnomelane is present only where the rocks have been metamorphosed by later igneous intrusions, and minnesotaite is rare. In the silicate rocks of the Mesabi range, according to Gruner (39, p. 27), "Nearly always two and commonly all three silicates, minnesotaite, stilpnomelane, and greenalite, are present." Gruner considers all three minerals as

primary, derived from colloidal gels of variable composition. As seen under the microscope, the textures of these rocks give unmistakable evidence that the present crystal form of much of the minnesotaite and stilpnomelane is post-sedimentary in origin (see, for example, plates II and III in Gruner's report). Quite possibly this is attributable to mineral growth during lithification, although the general absence of minnesotaite and stilpnomelane in the stratigraphically equivalent iron-formation of the Gunflint range would suggest otherwise. In the other ranges, greenalite is absent or rare. In some of the

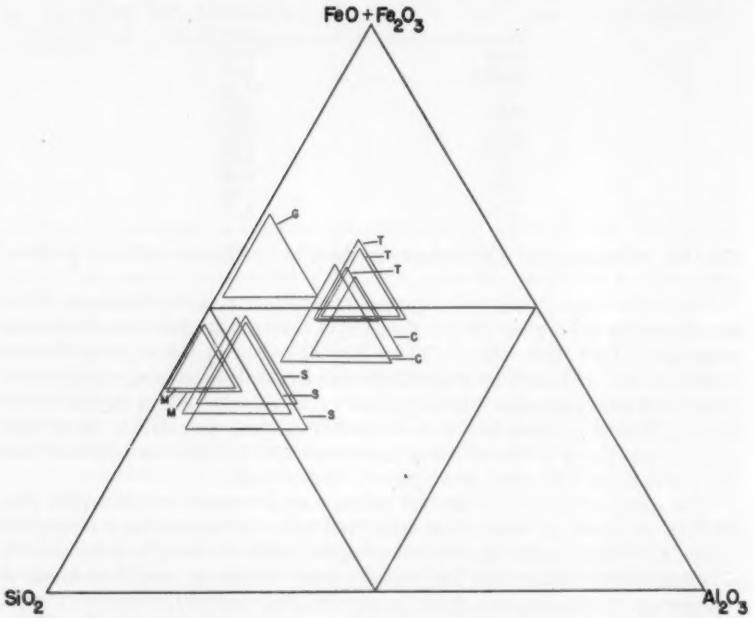

FIG. 4. Typical analyses of some of the common hydrous iron-rich silicates, in terms of total iron oxides, silica, and alumina. Each triangle represents a single analysis; area within triangle indicates total of constituents other than iron oxides, silica, and alumina. Weight percent. G: Greenalite; M: minnesotaite; S: stilpnomelane; T: thuringite; C: chamosite.

rocks, minnesotaite is the common silicate, in others stilpnomelane, and in some specimens the two minerals form alternate layers as shown in Figure 31. The issue resolves to this: If stilpnomelane and minnesotaite are secondary minerals, as might be concluded from textural relations, then one must assume that in some rocks the primary material was of exactly the right composition as to yield, upon slight metamorphism, layers of virtually pure stilpnomelane or minnesotaite. The writer is inclined to favor, with reservations, the view

that the two minerals are derived by metamorphism from preexisting silicate material of variable composition.

With regard to chlorite, again the textural relations to other minerals indicate a secondary (metamorphic) origin, but, inasmuch as chloritic layers in iron-formation almost always contain magnetite, carbonate, and chert, one can accept the metamorphic origin without difficulty; it is the monomineralic layers that cast the most serious doubt on metamorphic origin of stilpnomelane and minnesotaite.

Occurrence

As stated previously, granule-bearing silicate iron-formation is a major primary facies of the Biwabik iron-formation of the Mesabi range (61, 39). In association with magnetite-banded rock it recurs throughout the entire thickness (350–750 feet) and predominates over the hematitic and carbonate varieties that are common at some horizons. Silicate iron-formation, also of the granule-bearing type, is the dominant type in the Gunflint range (29).

Thin-banded nongranular silicate or silicate-carbonate iron-formation is the major rock type of the Cuyuna range. In the Gogebic range, silicate is present in most of the iron-formation and is abundant in the "slaty" members. In the Marquette range, silicate is a dominant constituent of the middle part of the Negaunee iron-formation between the upper oxide-rich part and the lower carbonate-rich part. Nongranular silicate iron-formation forms a separate stratigraphic unit (the "magnetic ironstone") in the Iron River-Crystal Falls district (50).

Description

At least two major types of silicate iron-formation can be distinguished. The first is exemplified by the granule-bearing greenalite rocks of the Mesabi and Gunflint ranges. These rocks are described by Leith (61, p. 101–102) as ". . . dull, dark green rocks of rather uniform fine grain and with conchoidal fracture. . . . On close examination, and particularly when the surface is wet, there may be observed numerous ellipsoidal granules of a green substance very slightly lighter green than the matrix in which they lie." The bedding is marked by layers of chert, carbonate, or iron oxides. Gruner has since shown (39) that minnesotaite and stilpnomelane are major constituents along with the greenalite in these rocks. The granules that characterize the rock are typically about 0.5 mm in diameter and irregular or rudely elliptical in cross section. They may lie in a matrix of chert or of other silicate material. Examples of greenalite structure are shown in Figures 25–27, which are photomicrographs of rocks from the Gunflint range. The granules lack concentric structure but commonly show an obscure mottling that in some specimens is seen to consist of tiny "spherites" of nearly circular cross section. In rocks from the Gunflint range, some of the spherites within the granules consist merely of tightly packed rounded areas of greenalite with indefinite borders in a "matrix" of greenalite and chert; others show a core of greenalite and an outer rim of radially arranged chalcedonic silica. Allowing for variations that are due to the particular spherical sections cut, the spherites are fairly uniform

in size, typically measuring about 0.05 mm in diameter. Structures of identical size and occurrence are described by Gruner (39) in Mesabi range rocks. In the material illustrated by Gruner, some of the spherites are entirely opaque, others are cored by an opaque substance. In the caption to his Plate VB, Gruner states: "Dusky dark rings and their centers probably are graphite although some may be magnetite. They may be of organic origin." Spencer and Percival (83, p. 375) describe similar structures in nongranular banded hematite jasper from Singhbhum, India, as follows: "Where well developed they consist of a central core or spherule of dense iron oxide with a surrounding clear 'halo' of crystalline silica. . . ." They conclude (p. 379): "We have seen no evidence whatever of microorganisms in the Indian banded jaspers, and failing direct evidence to the contrary, we would suggest that all these micro-spherulitic structures have been formed by crystallization from a silica-iron oxide gel or colloid." The conclusion reached by Spencer and Percival seems equally applicable to the spherites described in greenalite granules of the Gunflint and possibly to those described by Gruner from the Mesabi.

Granule-bearing silicate rock is found in the middle part of the Negaunee iron-formation of the Marquette range; many of the silicate layers, which consist principally of minnesotaite, show the typical granule structures. The rock is rich in magnetite, which forms layers up to an inch in thickness, and in some layers, magnetite is the sole constituent of the granules, as shown in Figure 28. Silicate granule rocks do not appear to be common in the Ironwood iron-formation of the Gogebic range, although some stilpnomelane-rich beds in the Tyler Fork section show what may have been granule texture before metamorphism (Fig. 29). In some specimens, textural relations suggest that granule-like areas of magnetite, carbonate, and chert may have been derived by diagenetic replacement of original silicate. Recognizable clastic material appears to be generally absent in these silicate rocks of the Mesabi, Gunflint, and Marquette districts, and the conclusion seems justified that the rock is the product of direct chemical or colloidal precipitation. The granules probably originated by slight current or wave action on a fine-grained precipitate, but the irregular outlines of the granules indicate that transport was slight.

The second major type of silicate iron-formation is nongranular and typically thin-bedded or laminated. This rock is associated with either carbonate-rich or magnetite-rich types. The specimen shown in Figure 18 is light green and laminated; it consists chiefly of minnesotaite with magnetite and carbonate in some layers. In the Crystal Falls district the principal iron-formation consists of banded chert and carbonate, but a silicate facies is present in some parts of the unit. Figure 20 shows the appearance in outcrop of an exceptionally thick bed of silicate; the silicate rock is dark greenish brown, laminated, and composed almost entirely of stilpnomelane. A photomicrograph of a thin laminae of fine-grained, brownish stilpnomelane is shown in Figure 32. Pyrite, apparently primary in origin, is a common accessory, forming layers or lenses. The stilpnomelane present as layers is dense and practically isotropic in thin section because of the fine grain, but in fine powders or in disseminated grains it is seen to be in definite crystal units. Figure 5 is a

sketch of two thin sections taken from a single specimen, which indicates the complex character of this iron-formation. The variations in the $FeO:Fe_2O_3$ ratio of the stilpnomelane, as indicated by gamma index values, offer strong support to the idea that the present composition of the layers is essentially that of the primary rock. Certainly if the stilpnomelane were the product of reaction between carbonate and chert, equilibrium would be attained through greater thickness of rock; actually equilibrium seems to have been attained only within units on the order of a few millimeters in thickness.

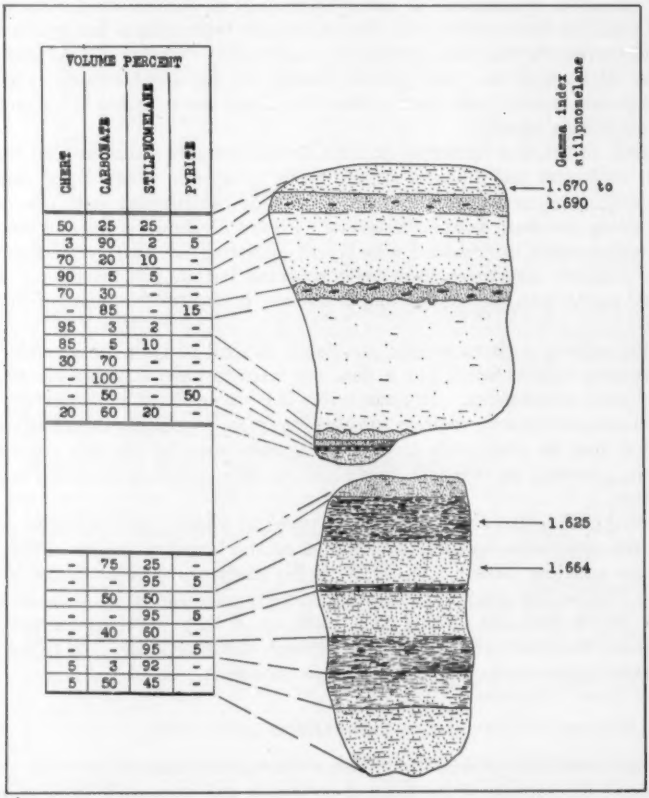

FIG. 5. Sketch of two thin sections from single specimen of iron-formation from Paint River dam outcrop, Crystal Falls, Mich. Illustrates complex character of rock. Note layer-to-layer variations in stilpnomelane index; in uppermost index determination, range in value may be due to contamination from adjacent layer, rather than variation within layer indicated. Note stylolitic contacts in upper specimen.

Silicate iron-formation in which clastic material is present as scattered grains or as thin layers is a common rock type. A specimen of this material is shown in Figure 19, which is a photograph of the "magnetic ironstone" from the Crystal Falls district. Similar rock was described from the Iron River district (50). The ironstone is a flinty laminated rock in which the layers are made up of mixtures, in varying proportions, of iron-rich chlorite, iron carbonate, chert, and magnetite. Stilpnomelane is locally present, and is a major constituent in some of the associated clastic beds. Stilpnomelane is also found as coarse-grained selvages to thin pegmatite veins that cut the rock (3). Because of affinities of this silicate rock to definite clastic rocks, the conclusion has been drawn (50) that it actually represents a fine-grained mechanical sediment that was enriched in iron during deposition. As shown in Figure 30, some of the "slaty iron-formation" of the Gogebic range is a laminated or thin-bedded rock that contains fine clastic material in a stilpnomelane-chert-carbonate matrix.

Much of the iron-formation in the Cuyuna range is a thin-bedded rock in which carbonate and silicates appear to be present in about equal amount. Both stilpnomelane and minnesotaite are major constituents, with one or the other being dominant in specific layers. Figure 31 shows a layer of virtually pure minnesotaite interbedded with layers consisting dominantly of stilpnomelane. Granule structures have been reported but in general appear to be scarce; clastic material, though not abundant, is observed in some of the material.

Magnetite is a characteristic associated mineral of both the granular and nongranular silicate facies, but it does not seem to have a systematic relation to the other constituents. In some layers it is exceedingly abundant and may be a major constituent, whereas in other layers of practically the same mineralogy, it may be completely absent. Magnetite may be the sole constituent of some granules, as shown in Figure 28, or may occur as scattered crystals in others.

The granules that characterize some types of silicate iron-formation do not show the concentric layering so common to the hematite ooliths. This fact has been noted by Gruner (34, p. 58) in his study of the rocks of the Mesabi range. Either the original structures were granules similar to those of glauconite, which does not form a true oolith, or, if they were similar to oolitic chamosite, the structure has been lost through metamorphism. In either case, the present granules would lack concentric structure.

Composition

The composition of the silicate-rich rocks varies within wide limits, chiefly because of the varying proportions of carbonate and iron oxide present in addition to the silicate. Not all available analyses are listed, because of uncertainty regarding analytical procedure or mineralogical composition, but those given provide a fair idea of the chemical composition.

The variations in composition of the rocks from the Mesabi range (analyses 1 to 5) are due chiefly to variations in percentage of chert and iron carbonate;

otherwise they show a general uniformity, especially in the low content of Al_2O_3 and in the $FeO:Fe_2O_3$ ratio. Actually, a complete gradation could be obtained from the Mesabi between rock composed of chert and carbonate and rock composed of chert and silicate.

TABLE 8

ANALYSES OF SILICATE IRON-FORMATION

	(1)	(2)	(3)	(4)	(5)	(6)	(7)	(8)	(9)	(10)
SiO_2	50.96	51.54	42.48	65.42	49.43[d]	36.40	52.85	51.18	48.11	34.18
Al_2O_3	1.09	0.08	0.53	0.08	none	3.54	8.71	11.95	3.27	15.04
Fe_2O_3	5.01	5.29	5.63	4.19	8.16	14.07	24.03[a]	8.09	13.62	10.89
FeO	30.37	26.51	33.76	23.63	30.52	20.93	—[c]	12.15	16.69	15.93
MgO	5.26	4.22	4.29	2.45	4.77	4.25	2.87	2.42	2.91	1.65
CaO	0.04	n.d.	n.d.	n.d.	0.18	3.50	0.10	1.12	0.80	1.00
Na_2O	none	n.d.	n.d.	n.d.	none	n.d.	1.48	2.12	0.24	0.11
K_2O	none	n.d.	n.d.	n.d.	none	n.d.	1.89	1.86	2.32	0.48
H_2O-	0.75	n.d.	n.d.	n.d.	0.51	n.d.	n.d.	0.07	0.44	8.92
H_2O+	6.41	3.20	4.80	4.16	6.24	1.57	5.56[b]	1.19	1.74	[c]
CO_2	none	7.34	7.44	n.d.	0.43	9.44	0.20	3.70	5.62	1.54
P_2O_5	none	n.d.	n.d.	n.d.	none	2.29	0.78	0.54	0.44	0.32
S	tr.	n.d.	n.d.	n.d.	tr.	n.d.	0.03	n.d.	n.d.	0.22
MnO	none	n.d.	n.d.	n.d.	0.17	1.95	1.10	2.71	3.27	tr.
TiO_2	n.d.	n.d.	n.d.	n.d.	none	0.03	0.60	0.51	0.52	0.79
C	0.21	n.d.	n.d.	n.d.	n.d.	0.20	n.d.	n.d.	n.d.	n.d.
Total	100.10	98.18	98.93	99.93	100.41	98.17	100.20	99.61	99.99	
Fe+Mn	27.11	24.47	20.18	21.30	29.43	27.65	17.77	18.20	25.03	21.26

[a] All iron given as Fe_2O_3.
[b] Includes organic material.
[c] Not given, but probably about 9 percent.
[d] Divided into "soluble" 29.90, "insoluble" 19.53.

1. Greenalite rock, Mesabi range (Leith, 61, p. 108, specimen 4).
2. "Taconite consisting of about 20% quartz, 18.5% siderite, and 61.5% minnesotaite." From lower cherty division, Biwabik iron-formation, Mesabi range (Gruner, 39, pp. 56–57).
3. "Taconite consisting of 8.5% quartz, 19% carbonate, and 71% minnesotaite and greenalite." Lower slaty division, Biwabik iron-formation (Gruner, 39, pp. 56–57).
4. "Taconite consisting of 28% quartz and about 72% minnesotaite." From lower cherty division, Biwabik iron-formation (Gruner, 39, pp. 56–57).
5. "Greenalite rock." Mesabi range (Jolliffe, 54, p. 416). Contains other silicates, chert.
6. Silicate-carbonate iron-formation, Cuyuna range. Average of analyses 7 and 13 (Harder and Johnston, 43, p. 120).
7. Chlorite mudstone. Dunn area, Crystal Falls district. Analysis by J. G. Fairchild, U. S. Geological Survey.
8. "Magnetic ironstone"—laminated flinty rock composed of chlorite, magnetite, carbonate, and chert. Outcrops on Stambaugh Hill, Iron River district. Analysis by Leonard Shapiro, U. S. Geological Survey.
9. Laminated rock similar to No. 8. From Hiawatha mine drill hole, Iron River district (James, 50).
10. Chamositic oolite, from Northampton sand ironstone, England (Taylor, 86, p. 61). Given for comparison.

Analyses 6 to 9 show a much wider range in composition. The material represented by analyses 7, 8, and 9 has been studied by the writer, and the conclusion drawn that the rock is a fine clastic that was enriched with iron during deposition. The alumina content is generally high, as compared with

the granule-bearing Mesabi rocks. Analysis 10 is of a chamositic oolite from the Northampton ironstone of England. This rock contains no free silica, and the high content of Al_2O_3 provides a strong contrast with the Lake Superior rocks.

Distinctive Features

Silicate rock of the Mesabi type is distinguished by granule structures in some layers. In general the silicate rocks, even where unmetamorphosed, contain a large but variable amount of magnetite, more than any of the other facies of iron-formation except the magnetite-banded rocks.

Nongranular silicate and silicate-carbonate iron-formation rocks are characterized by fine banding and the absence of clean chert layers. In thin section, some rocks of this type contain recognizable clastic material.

SUMMARY OF THE FACIES CHARACTERISTICS AND RELATIONSHIPS TO DEPOSITIONAL ENVIRONMENT

The conclusions regarding general characteristics and relationships of the four sedimentary facies of iron-formation are summarized below:

The *sulfide* facies is represented in the Lake Superior region by pyritic black slates. The rock originated as a black organic mud in sea water abnormally rich in iron. The depositional environment was characterized by poor bottom ventilation, with oxygen not being sufficiently abundant to destroy the organic material showered down from the more highly aerated upper zones. Iron in solution in the sea water is precipitated as sulfide by the H_2S that is generated by bacterial action on the organic material. In part the sulfur is derived from organic proteins, but the major contribution is almost certainly derived by bacterial reduction of sulfates in the sea water.

The *carbonate* facies consists of interbedded chert and iron-rich carbonate in roughly equal proportions. This rock was deposited as a chemical precipitate under reducing conditions in which oxygen was abundant enough to remove most of the organic material, yet not sufficiently abundant to cause oxidation of the ferrous compounds. The carbonate lacks oolitic or granular structure; it appears to have accumulated as a fine mud below the level of wave action. The carbonate is a molecular mixture of four components—$FeCO_3$, $MgCO_3$, $MnCO_3$, and $CaCO_3$—with $FeCO_3$ normally accounting for 70 percent or more of the total.

The *oxide* facies contains two principal types: hematite-banded and magnetite-banded. The *hematite-banded* rocks consist of interbedded hematite and chert or jasper. The hematite is now crystalline to a degree dependent upon the metamorphism undergone; in areas of low-grade metamorphism it is now fine-grained, gray or bluish specularite. The hematite is believed to have been deposited as hydrated ferric oxide in shallow, well-aerated waters. Oolitic structure is present in many layers, the ooliths consisting of thin skins of hematite and silica in all proportions ranging from a mere film of hematite around a chert core to a solid granule of hematite. The ooliths are believed to originate by the rolling action of waves and currents on chemically deposited

material. Mechanical transport of ooliths prior to final entrapment accounts for the lack of homogeneity and composition of the ooliths; in many respects the oolitic rocks possess definite affinities with clastic deposits.

The *magnetite-banded* rocks consist of layers of magnetite alternating with layers that contain varying proportions of iron silicates, carbonate, and chert. On the basis of associated minerals such as carbonate and hematite, the conclusion that the rock originated in an environment that ranged from mildly oxidizing to mildly reducing seems valid. The magnetite may have been formed by the settling of ferric oxide that was precipitated in upper well-oxygenated layers of the water into bottom environments in which the oxidation potential was too low to permit the stable existence of hematite. The abundance of post-depositional magnetite and carbonate in these rocks suggests also that diagenetic activity was strong.

The *silicate* facies comprises two major types, granular and nongranular. Both commonly, though by no means invariably, contain abundant magnetite and carbonate. The silicate may be greenalite, minnesotaite, stilpnomelane, or chlorite; of these, only greenalite appears to be definitely primary. The low content of alumina and potash indicates that neither glauconite nor chamosite, typical minerals of younger ironstones, were important primary constituents. In the Mesabi and Gunflint ranges, the silicates are commonly in the form of rounded to irregularly shaped granules. The granules differ from hematite ooliths in not possessing concentric rings; they more closely resemble glauconite granules.

Iron silicate, apparently either of primary origin or derived from a preexisting silicate, is found in association with minerals as contrasting as pyrite and hematite, so it is evidently a stable mineral over a wide range of oxidation-reduction environments. However, the optimum conditions for precipitation of the silicates, to judge from the associated minerals, are in the range from mildly oxidizing to mildly reducing. The conditions that would permit precipitation of iron and silica as a single mineral rather than as separate phases are not clear, although mutual precipitation of two colloids (in contrast to separate precipitation of two minerals from solution) is an obvious possibility.

THE PROBLEM OF THE CHERT

The problems related to the origin of the chert in the iron-formation rocks can be touched upon only briefly in this paper. The literature dealing with chert is extensive and suggests that chert may be formed in a variety of ways. The evidence seems to indicate that the chert of the iron-formation is a primary precipitate rather than a product of diagenesis or later replacement. The more important lines of evidence that lead to this conclusion are tabulated as follows:

1. The amount of chert in a given bed of iron-formation remains almost constant over wide areas. Post-consolidation replacements would yield a much more irregular pattern of distribution.
2. The amount of chert in a given type of iron-formation is reasonably constant. For example, the amount of chert in carbonate iron-forma-

tion is 30 to 35 percent, whereas that in oxide iron-formation is about 40 percent. It seems, therefore, that the chert must be related to the primary process of sedimentation.

3. As shown in an earlier paper (50), veinlets of chert are cut by stylolites that were formed before the beds were structurally disturbed. The veinlets quite reasonably may be considered diagenetic in origin, but they cut—with sharp-walled contacts—finer-grained layers of chert.
4. Slump structures, almost certainly formed before lithification, have been observed by the author in many of the rocks (50). These slumps involve thin bands of chert, therefore the chert is pre-lithification in origin. Similarly, intraformational breccia contains sharp-edged fragments of banded chert.

The abundance of chert in the Precambrian iron-formations is a highly distinctive feature, and although it has been shown by Hayes (46), and more recently by Kelley (55), that chert is present as interstitial material in some of the younger oolitic ironstones, the banded chert appears to be virtually limited to iron-rich strata of the Precambrian. Sakamoto (78) postulates that seasonal changes will account for alternating deposition of iron and silica. According to this theory, iron is transported in acidic solutions during wet seasons and precipitated when these solutions are neutralized, whereas silica is transported in alkaline solutions and precipitated in an acid environment. The theory is an attractive one, although, as pointed out by Tyler and Twenhofel (93), it does not explain the limitation of banded cherty iron-formation to the Precambrian.

DIAGENETIC MODIFICATIONS

The term diagenesis encompasses those changes produced in a rock after deposition and prior to lithification (90, p. 108). A very large number of modifications of the original sediment properly are to be ascribed to diagenesis. For example, it is probable that the pyrite in the pyritic slates originally was in another form, and that the hematite in the oxide rocks was originally a hydrous oxide. For the sake of brevity this discussion is limited to those changes that involve replacement of one iron mineral by another of different type.

As has been pointed out, the major factor in formation of the iron minerals is the oxidation-reduction potential (redox potential); and in general the burial environment of fine-grained sediments appears to possess a lower redox potential than the depositional environment. This is attributable chiefly to the decomposition of mechanically included organic material and to the fact that the sediment is more or less isolated from the oxygen-bearing sea water. A present-day example of profound diagenesis is described by Brujewicz (8, 9), who studied bottom conditions in the Barents Sea, off the north shore of western Russia; he reports that the water is well oxygenated from surface to bottom and that oxides of iron and manganese accumulate in brown deposits containing a considerable amount of organic material but no carbonate. Below the depositional interface, the brown layer grades into gray mud, much higher

in ferrous iron and lower in organic carbon. The amount of mechanically included organic material in the original deposit is apparently more than adequate, upon decomposition, to entirely reduce the oxides contained in the muds. Zobell (100) found that negative redox potentials are typical in the finer grained bottom sediments off the coast of California and Lower California; he also attributes the low potential chiefly to included organic matter. Emery and Rittenberg (24), in their study of drill cores in modern sediments of California, show that in addition to a lowering of Eh with depth, a slight increase in pH is evident.

In most fine-grained sediments of normal composition, the amount of organic matter included appears adequate to reduce all or practically all of the iron from the ferric state. However, it must be remembered that in an ordinary sediment the amount of reducible material—chiefly oxides of iron and manganese—is relatively small. Zobell (100, p. 484) points out that both the intensity and capacity factors of a system must be considered. He states: "It should be emphasized that the redox potential is an intensity factor in the same sense that temperature and pH are intensity factors. The redox potential does not indicate the reducing or oxidizing capacity of a system any more than the temperature of a system indicates how many calories of heat it may contain. . . ." The iron-formation, containing 5 to 10 times the amount of iron of a normal sediment, should show the effects of a lower redox potential during diagenesis, but the total effect ordinarily would not be large because of the bulk of reducible material present.

The observed modifications of the iron-formations that may be reasonably attributed to diagenesis are of a nature and extent that would be expected. Primary oxide rocks and associated silicate show some replacement by carbonate; carbonate rocks show comparable replacement by pyrite. In magnetite-banded rocks from the Gogebic range, some layers contain ooliths that apparently were originally hematite. These ooliths have been replaced by magnetite and carbonate in such a fashion as to largely destroy the oolith, the outlines of which either may be preserved by disconnected areas of hematite, or, as shown in Figure 33, may be almost completely lost. Figure 21 shows a dark graphitic phase of carbonate iron-formation from the Crystal Falls district; it illustrates typical diagenetic replacement by pyrite. The pyrite is rather coarse grained; it occurs as replacements around chert nodules and lenses, as paper-thin laminae, and as clusters of grains. Whether deposition of pyrite on the borders of the chert is due to greater susceptibility of the chert to replacement or whether this surface simply represents a structural discontinuity along which solutions could move is not clear. Inasmuch as Emery and Rittenberg (24) have shown that increase of pH with depth of burial is common, it seems likely both factors are operative; the presence of conditions more alkaline than the depositional environment would favor replacement of the silica, and the structural discontinuity would permit passage of solutions.

In general the diagenetic minerals are considerably coarser in grain than the primary materials. Distribution is irregular, and the minerals typically form clots and veinlets.

RELATIONSHIP OF IRON-FORMATION TO VOLCANISM AND TO GEOSYNCLINAL DEVELOPMENT

The theory that deposition of iron-formation is related to volcanism was developed by Van Hise and Leith, who state (95, p. 516): "The iron salts may have been transferred from the igneous rocks to the sedimentary iron-formations partly by weathering when the igneous rocks were hot or cold, but the evidence suggests also that they were transferred partly by direct contribution of magmatic waters from the igneous rocks and perhaps in small part by direct reaction of the sea waters upon the hot lavas." This rather cautiously worded theory was accepted and considerably expanded by Collins, Quirke, and Thomson (15) to account for the iron-rich rocks of the Michipicoten ranges of Ontario, and by Aldrich (1) for the iron-formation of the Gogebic range. It was tentatively accepted by Wagner (96) for the banded ironstones of the Swaziland and Witwatersrand systems of South Africa, although no volcanic rocks are present. The concept has been accepted by many students of the iron-formations of India. Regarding the banded ores of the Noamundi area of India, Percival (71, p. 239) states that the banded iron-formation rocks ". . . are associated genetically with igneous rocks and the most probable source of the bulk of the ore and the silica was from hot magmatic waters and possibly also magmatic vapours, ejected under submarine conditions." Dunn (19, 20) also emphasizes volcanic affinities for other iron-formation rocks of India.

The major factor that led Van Hise and Leith to the volcanic theory was the apparent inadequacy of "ordinary weathering" to supply solutions of the proper type for precipitation of iron-formation. They based this belief chiefly on study of temperate-climate streams, which are exceedingly low in both iron and silica. However, as pointed out by Gruner (33), Gill (30), and James (50), under certain tropical or subtropical conditions the iron and silica content of stream water may be very high and entirely adequate to account for iron-formation deposition. Furthermore, material held in solution in sea water is not precipitated continuously, nor, in general, can a chemical precipitate be directly related to the local stream contributions to an open sea, as can the clastic fraction. The actual process of chemical sedimentation is highly complex and depends on the total effect of concentration, temperature, pressure, and other components of the system. The present accumulation of carbonate marl on the Bahama shelf in response to conditions that permit release of CO_2 is a well-known example of localized chemical (or biochemical) precipitation from sea water in which the Ca content differs but slightly from that in areas of non-deposition.

There is much evidence opposed to the volcanic theory. The correlation between volcanism and iron-formation during the Huronian of the Lake Superior region, stressed by Van Hise and Leith, is not close. In most of the ranges, no volcanic activity is indicated during the major iron-formation epochs. The great Negaunee iron-formation of the Marquette range, for example, is underlain by quartzite and dolomite that are separated by profound unconformity from the underlying Keewatin volcanics; above, the Negaunee

is separated by strong unconformity from the Clarksburg volcanics of the upper Huronian. A similar situation prevails in the Menominee range. Volcanic rocks are present in the eastern Gogebic range, but are limited to the upper part of the Ironwood formation. No volcanic rocks of an age reasonably contemporaneous with the iron-formation are known in the Cuyuna range, nor in the Iron River-Crystal Falls district. Grout (31) and Gruner (33, 34) have shown the lack of correlation of volcanic activity with deposition of the iron-formation of the Mesabi range; they point out that deposition of the great iron-formations of the Huronian series took place during a much later period than the volcanism of the region, which was in the pre-Huronian. Similarly, in other parts of the world, notably South Africa and South America, evidence to indicate contemporaneous volcanism during the iron-formation epochs is lacking. Certainly no volcanic activity was present during deposition of the great bulk of younger ironstones of Europe and the United States. It is true that in some regions certain cherty sediments are associated with submarine volcanics (17, 79, 70) and that iron-formation is interlayered with volcanic rocks in the Keewatin of Canada. But the total bulk of these materials that can be shown to be directly associated with and perhaps genetically related to volcanism is small when compared with the iron-formation not associated with volcanic activity.

It seems to the writer that volcanism, though not uncommon during the deposition of the major iron-formations, does not have a close enough correlation in space and time with the iron-rich sediments to be genetically related to those sediments. Some less direct relationship is indicated. The stratigraphic record of the Huronian series in Michigan, shown diagrammatically in Figure 6, provides a hint as to an alternate interpretation. The stratigraphic terms used will be those that apply to the Marquette district, which contains the most complete Huronian section of any of the American ranges. The lower Huronian comprises a basal quartzite, the Mesnard, which is a massive orthoquartzite showing abundant crossbedding and ripple marking, and a clean, massive dolomite, the Kona. Algal structures, crossbedding, ripple marks, and mud cracks are common in the dolomite. Both the Kona and the Mesnard are typical "stable-shelf" or "foreland" types deposited in shallow waters under stable orogenic conditions.

The rocks of the upper Huronian are in striking contrast to those of the lower Huronian. The basal strata, the Goodrich quartzite, consist of conglomerate and dark quartzite. These strata grade upward into slate with intercalated thin beds of iron-formation (Bijiki, Greenwood) and volcanics (Clarksburg). This suite is overlain by the main part of the Michigamme formation, which consists of graywacke, slate, and basic volcanics, with an aggregate thickness probably much greater than that of all the underlying strata of the Huronian series. These rocks are of the so-called "geosynclinal" type, deposited in a strongly orogenic environment.

The rocks of the middle Huronian, lying between these contrasting sedimentary types, consist principally of the Negaunee-iron-formation, which locally attains a thickness of nearly 2,000 feet. Virtually a complete sedimentary cycle is represented. The lowest unit, separated from the lower Huronian by

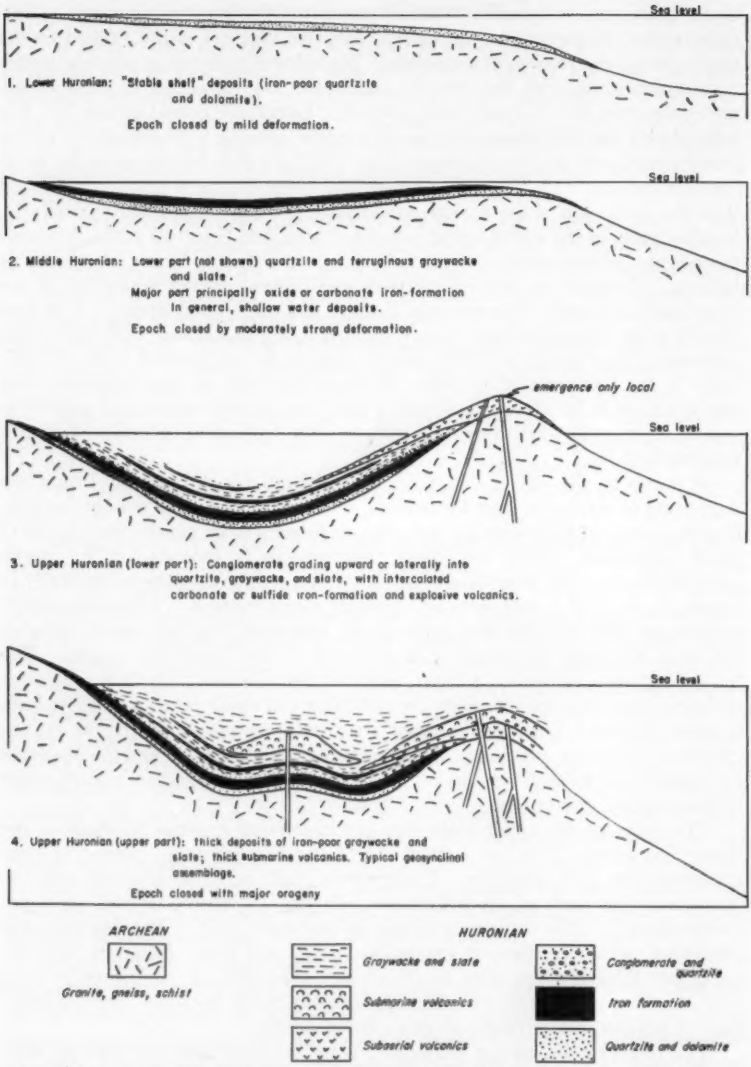
Sea level
1. Lower Huronian: "Stable shelf" deposits (iron-poor quartzite and dolomite).
Epoch closed by mild deformation.
Sea level
2. Middle Huronian: Lower part (not shown) quartzite and ferruginous graywacke and slate.
Major part principally oxide or carbonate iron-formation in general, shallow water deposits.
Epoch closed by moderately strong deformation.
emergence only local
Sea level
3. Upper Huronian (lower part): Conglomerate grading upward or laterally into quartzite, graywacke, and slate, with intercalated carbonate or sulfide iron-formation and explosive volcanics.
Sea level
4. Upper Huronian (upper part): thick deposits of iron-poor graywacke and slate; thick submarine volcanics. Typical geosynclinal assemblage.
Epoch closed with major orogeny
ARCHEAN
Granite, gneiss, schist
HURONIAN
Graywacke and slate
Submarine volcanics
Subaerial volcanics
Conglomerate and quartzite
Iron formation
Quartzite and dolomite

FIG. 6. Diagrammatic outline of Huronian depositional history in Michigan.

slight unconformity, is the Ajibik quartzite, which is overlain by the Siamo slate, much of which is iron-rich. Clastic sedimentation gradually ceased, and the uppermost part of the Siamo slate grades into the chemical deposits of the Negaunee iron-formation with interbedding at the contact. The character of the Negaunee itself reflects gradual shallowing of the basin of deposition, the lower part being of the carbonate facies and the upper part of the oxide facies. The epoch of iron sedimentation was terminated by structural disturbance.

The deposits of the lower Huronian are greatly different from those of the upper Huronian, but both are standard lithologic assemblages that may be found in systems of every major era. They have one notable feature in common: both contain iron in only normal amounts. Deposition of the iron-rich beds in the middle Huronian and, to a much lesser extent, in the lower part of the upper Huronian, suggests that some special conditions prevailed during this particular period of time. The environmental requirement for deposition of iron-formation, as discussed earlier in this paper, is the restricted basin. This requirement coincides in time with what would be a normal stage of development of the Huronian geosyncline in the region—that is, the formation of a low, offshore buckle or swell. Insofar as sedimentation is concerned, this may be termed the "critical period" in the history of a geosyncline. In this initial stage in the orogenic cycle, the effect on the land area is relatively slight, especially if the land area is in an advanced stage of peneplanation; the clastic load may briefly increase, then decrease as stream equilibrium is re-established. In the basin of deposition, however, the presence of a barrier to restrict circulation would permit development of abnormal concentrations in the water and abnormal redox potentials in the bottom environments. The environment is appropriate for deposition of chemical sediments, the composition of which will be governed by the nature of the land area, the climatic conditions prevailing, and the configuration of the basin.

In the later stages of geosynclinal development, the offshore buckle would be intensified and gradually develop into a volcanic chain analogous to those associated with island arc structures. Collapse of this belt in the advanced stages of the Huronian geosyncline is indicated by the deposition of the iron-poor beds of the upper Huronian.

On the basis of the evidence outlined, the conclusion is drawn that the relationship between volcanism and iron-formation deposition is structural, not chemical. In a sense, the association is accidental; both are related to the geosynclinal development. The structural position of the iron-rich strata has been noted by Cady, McKelvey, and Wells (10), who class the Precambrian iron-formations as characteristic of "eugeosynclines" and those of the Clinton type as characteristic of "miogeosynclines." But development of a geosynclinal basin is by no means a necessary factor in precipitation of iron-rich beds. Any combination of circumstances that will yield a basin with restricted circulation, which is bordered by low-lying land being weathered under appropriate climatic conditions, would suffice. The Mesozoic ironstones of Europe, according to students such as Déverin (16), were deposited in seas that were gradually advancing over land surfaces of low relief. Under these circum-

stances a deposit is possible whenever the sea gains access to a topographic basin that will be separated by a threshold from the open ocean. The thin and discontinuous lenses of iron-formation so common in the Keewatin volcanics of Canada conceivably may be due to deposition in small basins ponded by submarine volcanic accumulations.

The general setting in which deposition of the Huronian iron-formation occurred is summed up as follows:

1. During the early part of the Huronian, the land area was one of very low relief, shelving off gradually to the open ocean. The rocks of the land area were undergoing deep chemical weathering. The sediments deposited on the broad shallow shelves consisted of clean, washed quartz sand and of dolomite.

2. Slight structural disturbance at the end of the lower Huronian resulted in the development of broad offshore buckles in a belt peripheral to the land area. The restrictions imposed on circulation with the open ocean permitted abnormalities of concentration and bottom conditions in the marginal basins thus formed. In these marginal basins were deposited the chemical products of the deep weathering of the land area—chiefly iron and silica.

3. The marginal basins may have been and probably were relatively small in size. For example, the Gogebic range is directly on the strike projection of the Marquette range and lies only 65 miles to the west, yet dissimilarities in iron-formation stratigraphy indicate deposition in separated basins. Marginal basins on the order of 500 miles in length and 100 miles in width would be adequate to account for the present distribution of the ranges. The implications of this statement are that elsewhere in the region, where the bordering seas were not restricted by offshore buckling, normal sedimentation would prevail. The absence of iron-formation in the type Huronian section north of Lake Huron might appear to support this contention, though other possibilities, such as questionable correlation of Huronian strata, may account for the fact equally well.

4. The epoch of iron deposition was ended in the region by strong structural disturbances that produced profound changes in both the land area and the sea-bottom configuration. Chemical sedimentation gave way abruptly to clastic deposition, and the broad swells that earlier acted as barriers for shallow seas developed into a chain of volcanic islands.

The concept of deposition in restricted basins here expressed is very similar to that outlined by Woolnough (98), who suggests that many of the unusual chemical deposits were precipitated in "barred basins." Woolnough also stresses, by analogy with conditions prevailing in Australia, the possible great importance during long periods of geologic history of deep chemical weathering of an intensity not now observable on the present surface of the earth (97, 99). This combination of conditions—namely, the barred or restricted basin of deposition coincident with deep chemical weathering of the land surface—appears necessary to explain deposition of the iron-formations. The difference between the iron-formations of Precambrian age and those of

younger age is considered due to the intensity and duration of the deep chemical weathering; it is a difference in degree rather than kind.

U. S. GEOLOGICAL SURVEY,
IRON RIVER, MICHIGAN,
Oct. 1, 1953.

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Explanations of Figs. 7–33

Fig. 7. Banded pyritic slate, Iron River district. Light-colored layers rich in fine-grained pyrite; rock contains about 40 percent pyrite. Polished surface, natural size.

Fig. 8. Photomicrograph showing appearance of pyritic slate in polished section. Note crystal outlines of grains. × 400.

Fig. 9. Banded rock in which layers of carbonate (Cb) alternate with layers of pyrite (Py). Deposited at environment near boundary between carbonate and sulfide depositional zones. Polished surface, natural size. Iron River district.

Fig. 10. Banded chert-carbonate from Spies mine, Iron River district. Polished surface, etched with HCl. Carbonate gray, chert dark. Layering thicker than normal. Width of specimen about 7 inches.

Fig. 11. Thin-banded chert-carbonate from east end Gogebic range. Drill core. Chert white due to surface abrasion; carbonate gray. Note stylolite seam. Length of specimen about 5 inches.

Fig. 12. Specimen from James mine, Iron River district. Shows layer of hard blue hematite (dark) interlayered with carbonate and chert (both light colored). Polished surface. Width of specimen about 5 inches.

Fig. 13. Jaspilite in outcrop at Jasper Knob, Ishpeming (Marquette range). Dark layers are red jasper; light layers are silvery hematite. Width of photograph area, about 3 feet.

Fig. 14. Layers of limestone (white) interbedded with specular hematite (dark). Irregular white patches consist of secondary carbonate. From Curry member of the Vulcan iron-formation, Millie pit, Iron Mountain. Polished surface. Width of specimen about 6 inches.

Fig. 15. "Wavy-bedded" iron-formation, from dump of old Tyler mine, Gogebic range (Wisconsin). Magnetite (dark gray) in irregular layers separated by layers of chert-minnesotaite-carbonate (speckled). Polished surface, etched with HCl. Width of specimen, 5 inches.

Fig. 16. Even-bedded magnetite iron-formation, from dump of Vicar mine, east Gogebic range. Magnetite (dark gray to black) in even layers separated by layers consisting of greenish minnesotaite, chert, and iron carbonate. Polished surface. Width of specimen, about 5 inches.

Fig. 17. Laminated magnetite-stilpnomelane rock. Vicar mine, Gogebic range. Magnetite-rich layers light gray, stilpnomelane-rich layers dark. Speckled layer in upper part of specimen contains disseminated carbonate rhombs. Layer at top edge is minnesotaite. Dark veinlets are stilpnomelane mainly. Polished surface, etched with HCl. Width of specimen, about 4 inches.

FIG. 18. Laminated minnesotaite rock. Drill core from Gogebic range. Chiefly composed of pale green minnesotaite, with some magnetite-rich layers (dark). Width of specimen, about 2 inches.

FIG. 19. "Magnetic ironstone," Crystal Falls district. Light-colored layers are mixtures of chert and carbonate, gray layers carbonate with silicate (iron-rich chlorite), dark layers chlorite and magnetite. Specimen marginally oxidized. Polished surface. Width of specimen about 2½ inches.

FIG. 20. Photograph of stilpnomelane beds in Paint River dam outcrop. Crystal Falls, Michigan. Pocket knife gives scale. Stilpnomelane beds have distinctive coal-like cross-fracture, are interbedded with banded chert-carbonate rock. Photograph by F. J. Pettijohn.

FIG. 21. Dark carbonate rock with chert nodules, from Tobin mine, Crystal Falls, Mich. Shows development of diagenetic pyrite (white) along certain laminae and at boundaries of chert. Polished surface. Natural size.

FIG. 22. Hematite oolite, Warner mine, Amasa district, Michigan. Hematite ooliths (black) in chert matrix. Shows typical range in sizes, shapes, and constitution of the ooliths. Ordinary light, × 20.

FIG. 23. Moderately stretched hematite oolite, Michigan mine, Amasa district. Ordinary light, × 20.

FIG. 24. Deformed and partly recrystallized hematite oolite, Millie pit, Iron Mountain, Michigan. Flattened granules consist of finely specular hematite and chert, in chert matrix. Small amount of carbonate as small grains near center of photo. Ordinary light, × 30.

FIG. 25. Mixed oolite, Gunflint range, Ontario. Hematite ooliths show concentric rings separated by chert. Typical greenalite granule at right center. Ooliths at upper left and upper right composed largely of greenalite. Ordinary light, × 30.

FIG. 26. Greenalite granules in chert matrix. Gunflint range, Ontario. Typical. Ordinary light, × 30.

FIG. 27. Greenalite granules showing peculiar mottling of greenalite caused by recrystallization of chert. Gunflint range, Ontario. Ordinary light, × 30.

FIG. 28. Magnetite granules, in minnesotaite matrix. Middle part of Negaunee iron-formation, Marquette district. Ordinary light, × 20.

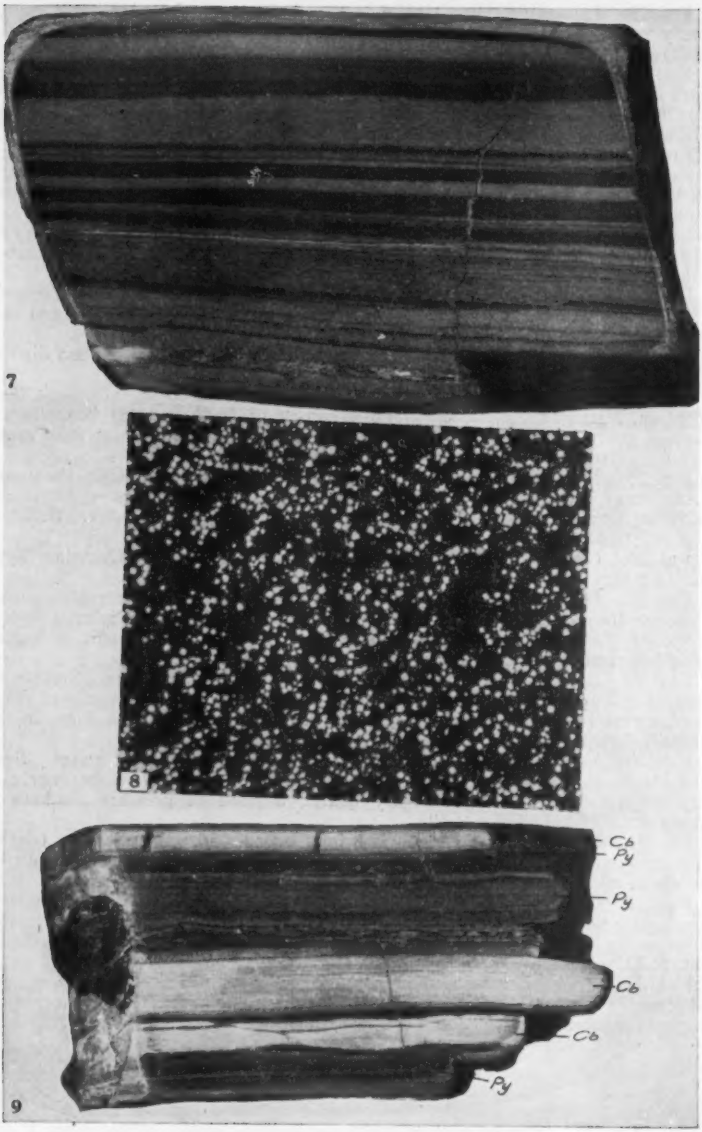
FIG. 29. Spotted rock, from Tyler Fork section, Gogebic range. Spots composed of stilpnomelane with scattering of magnetite crystals, in matrix of chert. Possibly represents metamorphosed granule rock such as shown in Figs. 26, 27. Ordinary light, × 50.

FIG. 30. Stilpnomelane "slate." Tyler Fork section, Gogebic range. Scattered clastic grains of quartz and abundant crystals of magnetite in matrix of platy stilpnomelane. Some secondary quartz, as that in pressure shadows of magnetite. Ordinary light, × 60.

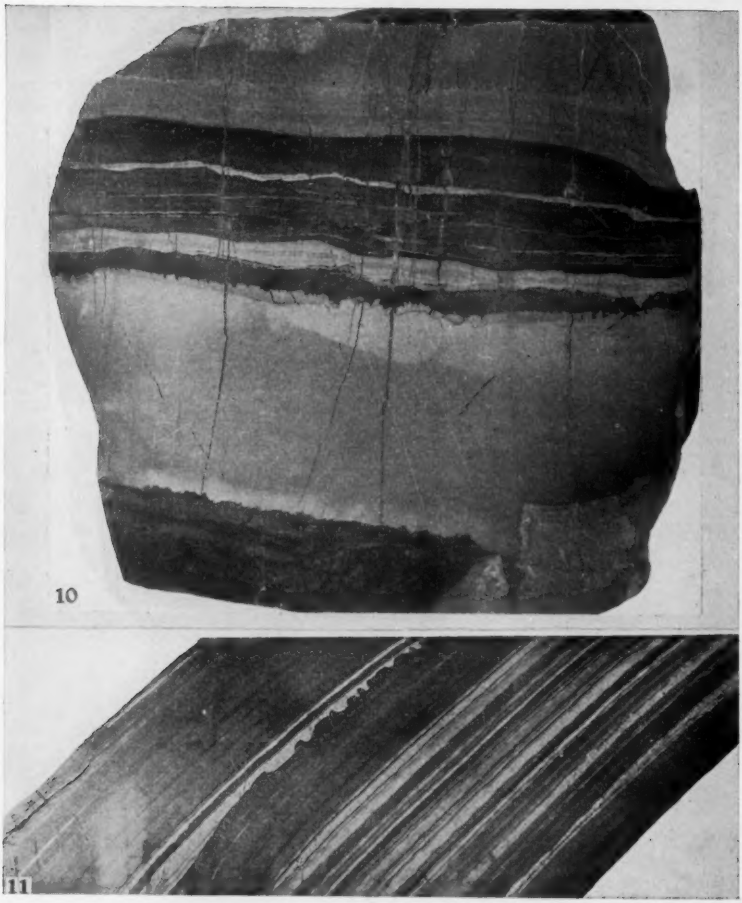
FIG. 31. Minnesotaite layer (light) interbedded with stilpnomelane (dark), with scattered grains of carbonate. Portsmouth pit, Cuyuna range. Needlelike crystals in minnesotaite layer also are minnesotaite. Ordinary light, × 33.

FIG. 32. Contact between layer of brown stilpnomelane (bottom) with layer of carbonate with scattered crystals of stilpnomelane. This contact is same as that shown on fig. 5 between the two top layers of the lower thin section. Ordinary light, × 30.

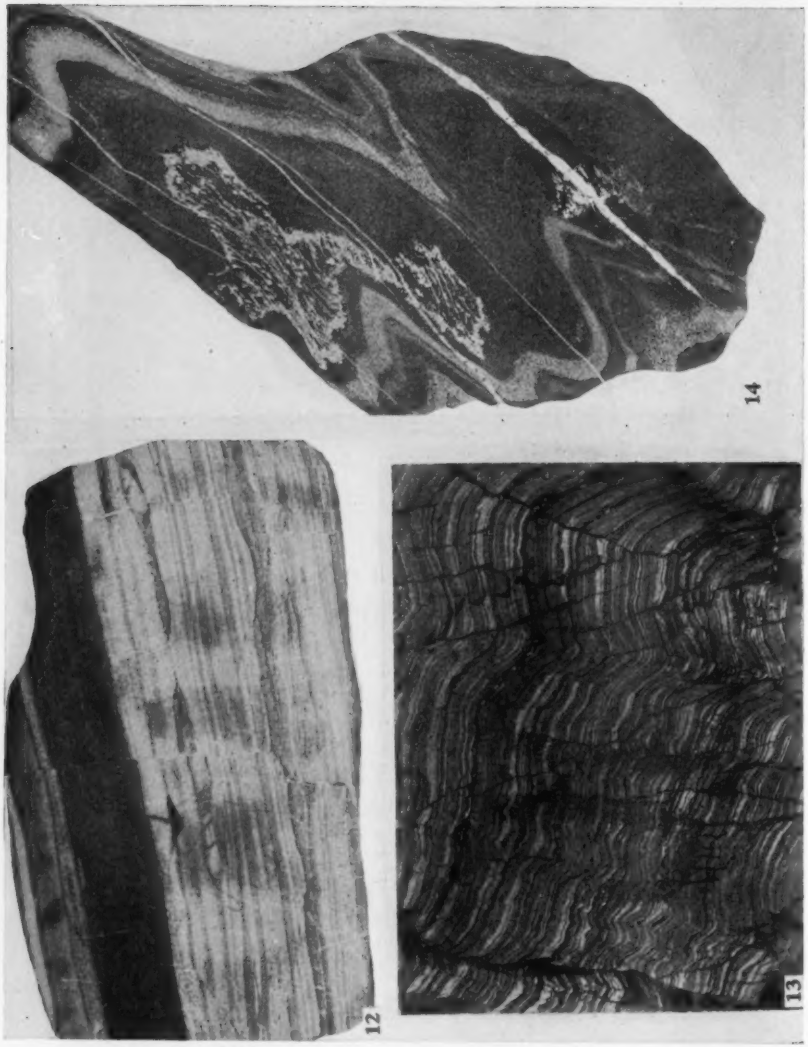
FIG. 33. Granules composed chiefly of coarse-grained iron carbonate, and magnetite. Chert matrix. Vicar mine, Gogebic range. Believed to represent complete diagenetic replacement of original hematite ooliths. Ordinary light, × 30.



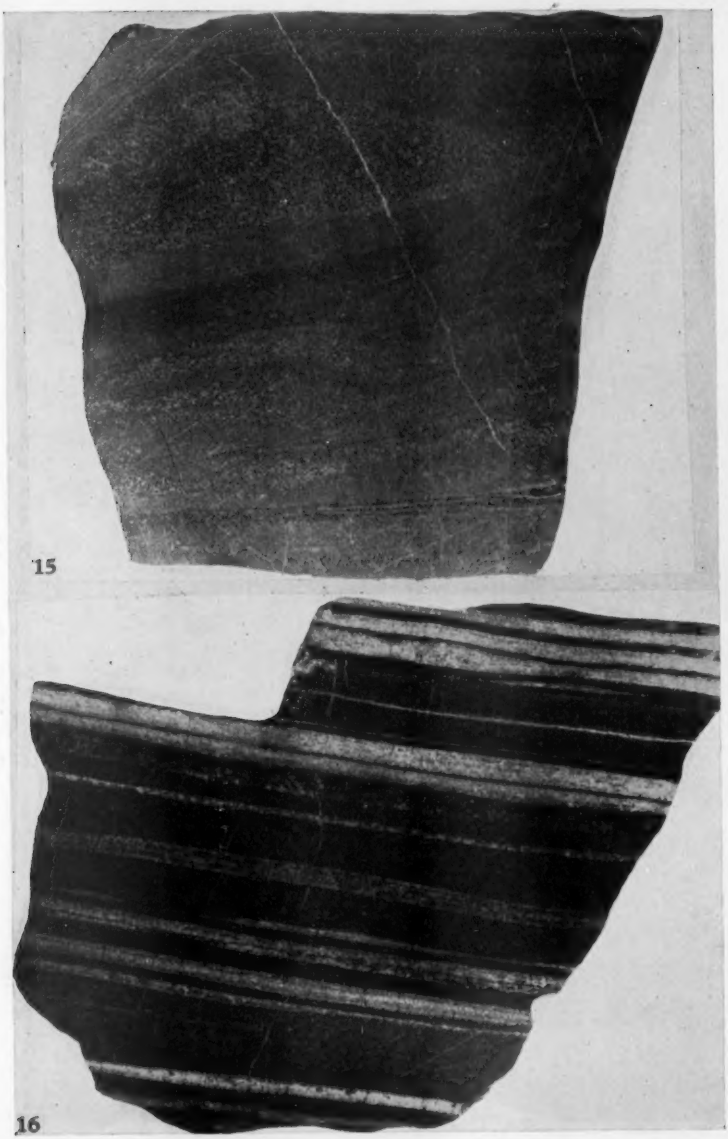
FIGS. 7–9. Pyrite-bearing rocks.



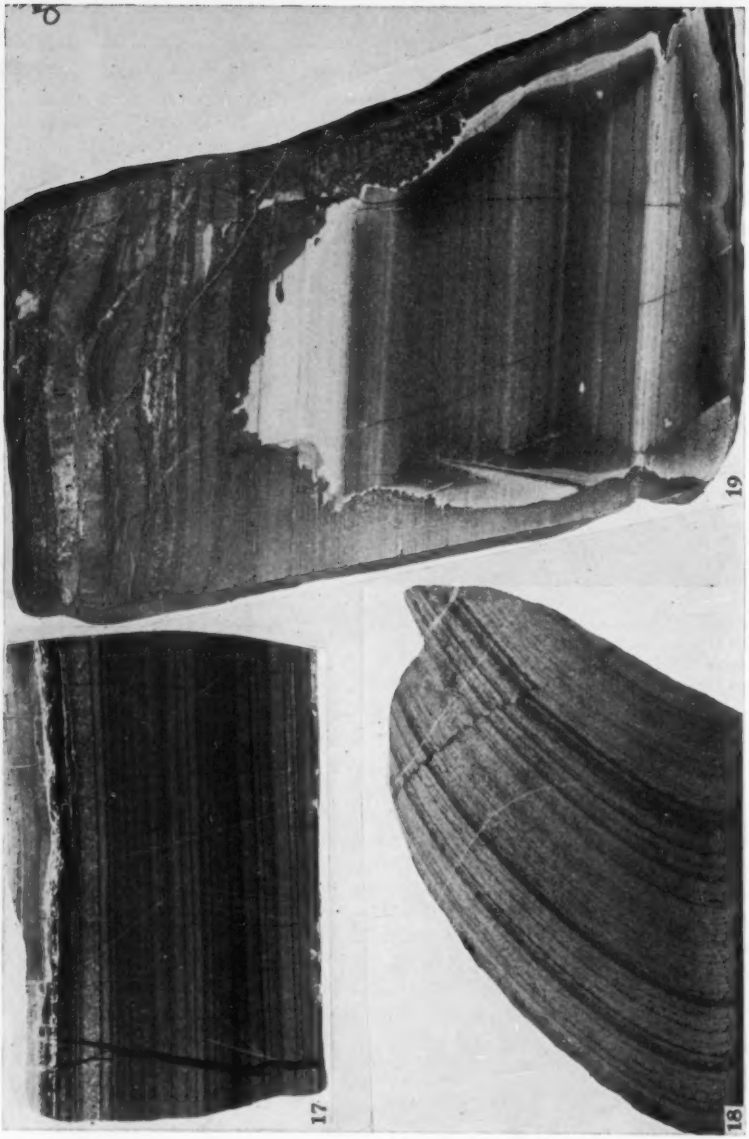
FIGS. 10–11. Examples of carbonate facies.



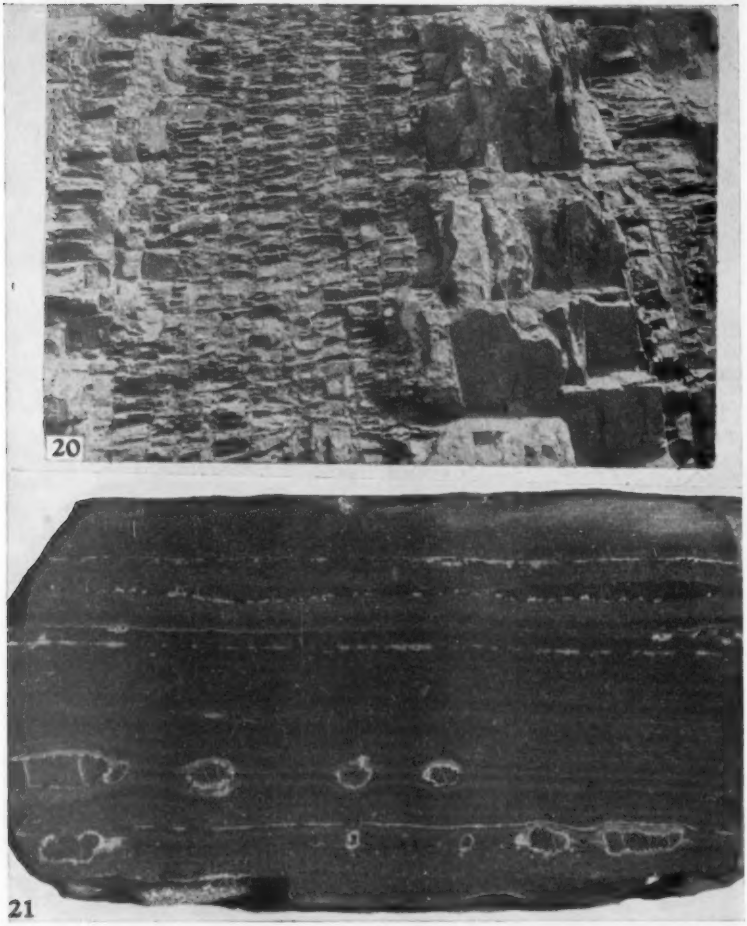
FIGS. 12–14. Iron-formation lithologies.



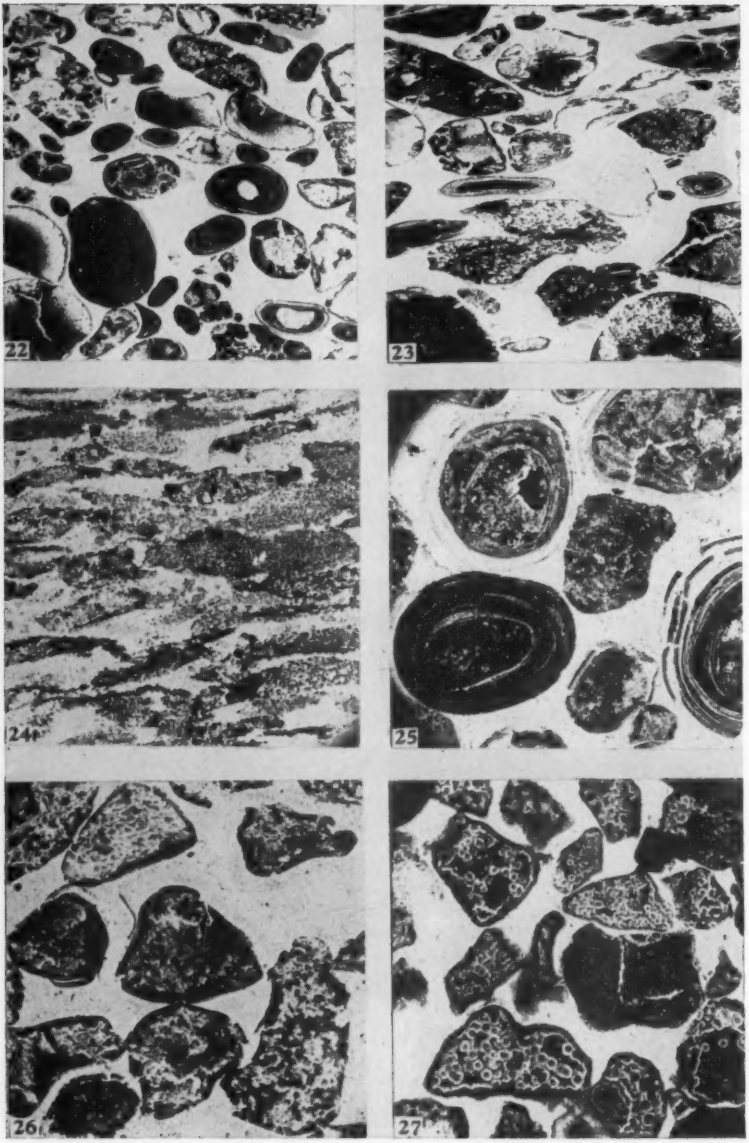
FIGS. 15–16.. Magnetite-banded iron-formation.



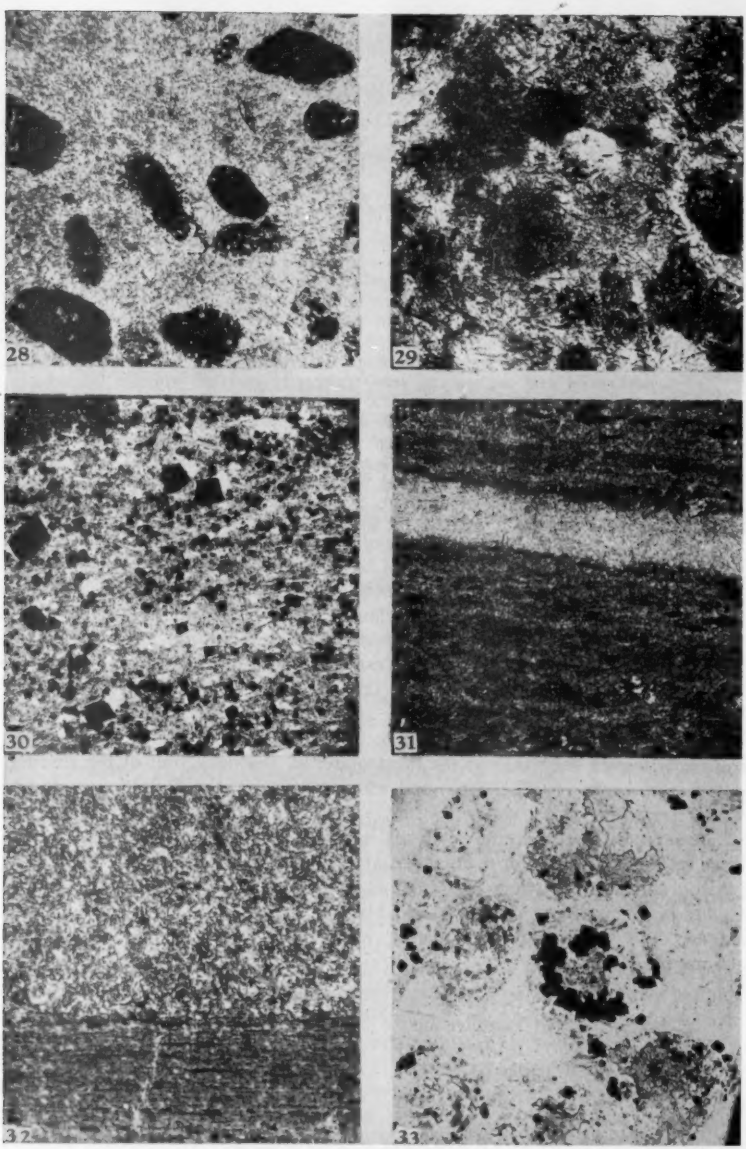
FIGS. 17–19. Silicate-rich iron-formation.



FIGS. 20–21. Iron-formation.



FIGS 22–27. Photomicrographs of iron-formation rocks.



FIGS. 28–33. Photomicrographs of iron-formation rocks.

INVESTIGATION OF RANK IN COAL BY DIFFERENTIAL THERMAL ANALYSIS[1]

HERBERT D. GLASS

ABSTRACT

Differential thermal analysis, using conventional equipment, has been used to differentiate between ranks of coal. Five distinct thermal curve types are recognized, based upon physical, chemical, and structural changes in coals of different ranks. Classification of curve type is based principally upon the number and temperature of endothermic devolatilization thermal peaks.

Natural rank boundaries, where curve types change, occur between meta-anthracite and anthracite, semi-anthracite and low-volatile bituminous, medium-volatile bituminous and high-volatile bituminous, and high-volatile bituminous and subbituminous. Gradational boundaries with no change in curve type occur between anthracite and semi-anthracite, and low-volatile and medium-volatile bituminous. The change in thermal curve type agrees favorably with A.S.T.M. rank boundaries.

Thermal curves indicate that the coalification process is gradual with increase in rank to anthracite. At this point, a structural "break" separates meta-anthracite from the rest of the sequence.

THE theory and technique of differential thermal analysis used in this study has been described by Grim and Rowland (1).[2] The method consists essentially of heating simultaneously the material to be analyzed and a thermally inert comparison material which does not undergo a phase or chemical change within the temperature range studied (room temperature to 1000° C.). The temperature is raised at a uniform rate and differences in temperature between the two materials are measured and recorded. When the temperature of the sample is higher than that of the inert material, the deflection of the thermal curve is upwards (exothermic), and when it is lower, the deflection is downwards (endothermic).

Differential thermal curves of coal were perhaps first attempted by Hollings and Cobb (2) using a nitrogen atmosphere and coke as the inert comparison material. In their experiments, loss of volatiles began at about 250° C, and between this temperature and 1100° C a number of fairly well defined stages could be distinguished during which exothermic or endothermic reactions predominated:

Endothermic	250°– 410° C
Exothermic	410°– 470° C
Endothermic	470°– 610° C
Exothermic	610°– 800° C
Endothermic	800°–1100° C

[1] Published by permission of the Chief, Illinois State Geological Survey

[2] Numbers in parentheses refer to References at end of paper.

These temperatures are essentially in agreement with those obtained by this author for high-volatile bituminous coals.

More recently, control of the pressure within the furnace has been used as a vacuum technique by Whitehead and Breger (3), Whitehead (4), and Breger and Whitehead (5). With regard to rank determinations, their results led to the conclusion that "there cannot be established a 'Standard Curve' for each type of coal with a definite peak or peaks—either exothermic or endothermic—characteristic of a given type of coal" (4).

X-ray and differential thermal analysis have been used in Japan to study the structural and chemical changes during coking (6, 7). The structure of coal is considered schematically as the "super-position of macromolecules of carbon forming aromatic nuclei to which are attached hydrocarbon rings and chains along the peripheral carbon atoms." This 2-dimensional lattice changes during the coking process to a 3-dimensional structure releasing the hydrocarbon molecules as gases.

Volatilization of hydrocarbon molecules results in an absorption of energy, causing endothermic reactions. Volatilization is accompanied by condensation of the carbon hexagonal planes, causing exothermic reactions. The loss of the primary coal structure occurs between about 230°–600° C (volatilization range). Above 500° C secondary thermal decomposition of the primary decomposition products occurs in bituminous coals with graphitization starting at about 800° C.

The coking process from 230°–800° C evidently consists of simultaneous endothermic and exothermic reactions, the thermal peaks representing the mean of opposing forces.

EXPERIMENTAL PROCEDURE

Conventional equipment was used and an inert atmosphere or vacuum was not required. A nickel sample holder gave better results than sulphur-resistant stainless steel. Frequent replacement was necessary, owing to formation of nickel sulfide. A loose-fitting nickel cover should be placed over the sample holder during the thermal run.

Curves were run from room temperature to 1000° C at a heating rate of about 10° C per minute. The sample size was about 0.5 gram. Packing of samples in holder was uniform with only a light amount of pressure. Excessive packing pressure prevented escape of volatile constituents, inhibited plasticity of the coal, and changed the shape of the thermal curve. All curves were run at 100 ohms resistance.

Differential thermal analysis of coal in air results in an exothermic oxidation curve; the rate of oxidation exceeds any absorption of energy resulting from devolatilization. If permitted, oxidation of coal will mask loss of volatile constituents. The placing of a loose-fitting cover on the sample block evidently inhibits the escape of gas from the coal and entry of air, and permits the building-up of volatile partial pressures. As gases escape almost continuously, no air can enter the system and the coal is constantly surrounded by escaping gases. The coal is, therefore, being heated in its own changing hydrocarbon atmosphere, and oxidation does not occur.

Under the experimental conditions described, the shape of the curve may depend on three principal factors: (1) the plastic state of the coal during devolatilization; (2) the types of organic chemical reactions which take place when coal is decomposed and volatile groups released; (3) the structural changes which occur during and following devolatilization. The relationships between the above factors furnish thermal criteria for rank determinations in coal.

THERMAL CURVES

Coal bed and locality are indicated on the thermal curves as well as the amount of volatile matter (V. M.) on the dry mineral-matter-free basis.

The principal features observed in thermal curves are: (1) the endothermic reaction accompanying dehydration between 120°–150° C; (2) the exothermic reaction which occurs between dehydration and the start of devolatilization (240°–610° C); (3) endothermic devolatilization peak or peaks between 435°–735° C, which vary in number, amplitude, and temperature, dependent on rank; (4) the sharp exothermic reaction at about 400° C characteristic of high-volatile bituminous coals, the cause of which is unknown; (5) the exothermic reaction following volatile loss caused by the final condensation of the hexagonal carbon planes and resulting in graphitization at about 800° C; (6) the endothermic reaction between 800°–1000° C, probably caused by evolution of hydrogen. The cause for reaction (2) is not apparent; it may well be caused by oxidation of the coal and is referred to as the low-temperature exothermic reaction.

The term plasticity is used as defined by Mott and Wheeler (8). "The plasticity of coal is a complex phenomenon which appears to be induced primarily by the pressure of gases causing surface flow at the moment when, under the action of heat, the molecules at the surface have attained a degree of fredom comparable with that obtaining in a liquid."

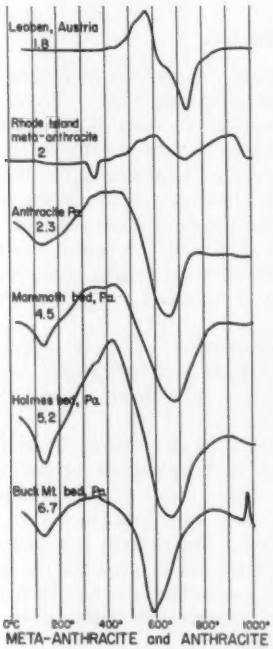
Meta-Anthracite

Two thermal curves are shown for meta-anthracite in Figure 1 (1.8 and 2 percent V.M.). The Rhode Island coal contains too much mineral matter (up to 40 percent) to give a satisfactory curve. The curves are characterized by a small endothermic reaction caused by volatile loss at the highest temperatures observed for any rank of coal (725°–735° C). The coals contain no moisture, and the low-temperature exothermic reaction which normally follows water loss also occurs at the highest temperature for any rank of coal (570–610° C). Volatilization reaches a maximum at the endothermic peak between 725–735° C.

The volatile loss peak at 735° C in meta-anthracite is much higher in temperature than the similar peak for anthracite; a higher temperature for the start of volatile loss and the lack of low-temperature water between 100–200° C also constitute important differences. The temperature for the volatile loss peak for anthracite (2.3 percent V. M., Fig. 1) is 660° C; a peak temperature difference of 75° C between coals with 1.8 percent and 2.3 percent

V.M. is noteworthy. The variation may be related to differences in the structural organization of the two ranks.

Although the curve for the Rhode Island coal is irregular owing to ash, the volatile loss peak at 725° C can be observed. The change in curve type from meta-anthracite to anthracite occurs between 2 and 2.3 percent V.M. This indicates that the A.S.T.M. rank boundary at 2 percent V.M. (9) is probably correct, and the great difference in peak temperature between the two ranks suggests that a "break" exists in the coalification process.



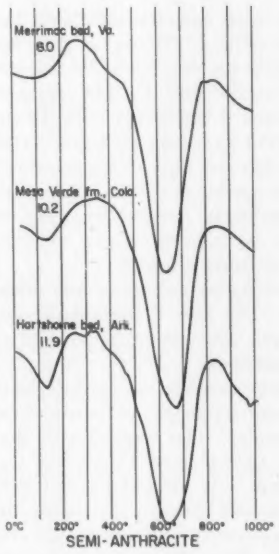


FIG. 1. (*Left*) Meta-anthracite and anthracite.
FIG. 2. (*Right*) Semi-anthracite.

Anthracite

Anthracite curves are shown in Figure 1 (2.3–6.7 percent V.M.). The features which distinguish anthracite curves are: (1) water loss between 125–140° C; (2) exothermic reaction between 420–470° C; (3) endothermic volatile loss peak between 660–680° C.

All reactions are lower in temperature than in meta-anthracite, and the size of the volatile loss peak is greater. The lower temperatures for the Buck Mountain coal are caused by kaolinite, whose endothermic dehydration peak

below 600° C evidently lowers the peak temperature. The exothermic peak at 980° C is also caused by kaolinite.

The characteristic feature of all meta-anthracite, anthracite, and semi-anthracite curves is the presence of a single volatile loss peak. Variation in peak temperature for the various anthracite coals may be caused by structure as well as impurities. The single volatile loss peak type curve at temperatures between 630–680° C (anthracite and semi-anthracite) is characteristic of all high rank coals which do not become plastic when heated, and is referred to as the anthracite type curve.

Semi-Anthracite

There is no essential difference between curves for anthracite and semi-anthracite (Fig. 2) except for peak size and temperature. Water-loss peaks are generally the same, but the low-temperature exothermic reaction now occurs between 250–350° C. The decrease in temperature of this reaction with decreasing rank is associated with the increase in amount of volatiles. As the V.M. content increases, the start of volatile loss occurs at a lower temperature causing a decrease in peak temperature for the exothermic reaction. The temperature range for the volatile loss peak (630–670° C) is also generally lower than for anthracite; its greater size is caused by the greater amount of volatiles lost. The decrease in volatile loss peak temperature with decreasing rank is evidently related to structure.

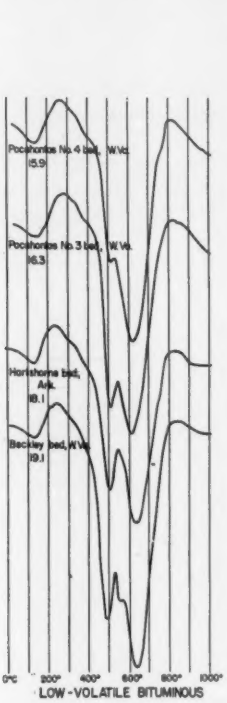
The curves for anthracite and semi-anthracite indicate a progressive rank sequence. The size of the volatile loss peak becomes greater, and all peak temperatures decrease progressively, as the rank decreases and the amount of V.M. increases.

The rank boundary with anthracite could not be precisely determined owing to the transitional nature of the curves and because too few samples were available for study. The transition to the larger curve of the Merrimac semi-anthracite (8.0 percent V.M., Fig. 2) from the Buck Mountain anthracite (6.7 percent V.M., Fig. 1) is not shown. The volatile loss peak evidently increases in size through this range, and the larger peak for the 8.0 percent V.M. semi-anthracite indicates that true semi-anthracite has been reached at this value. Thus the thermal analysis boundary probably agrees with the A.S.T.M. rank boundary at 8 percent V.M.

Bituminous Coals

Bituminous coals differ from coals of other ranks, among other characteristics, because of their plasticity. It has been shown that the non-plastic anthracite and semi-anthracite coals have a single volatile loss peak at temperatures between 630–680° C. On the other hand, all bituminous coals, low-volatile, medium-volatile, and high-volatile, can be classified into two types, each of which shows at least two volatile loss peaks. It seems evident that the occurrence of a second volatile loss peak only in bituminous plastic coals is not coincidental, and that the second peak must be related to the plastic nature of bituminous coals.

Low-Volatile Bituminous.—Thermal curves for low-volatile bituminous coals are shown in Figure 3. The plastic nature of the coals is indicated by the second thermal peak for volatile loss at about 500° C. The rank boundary between semi-anthracite and low-volatile bituminous is evidently a boundary of plastic capacity which should be revealed by the thermal curve. The



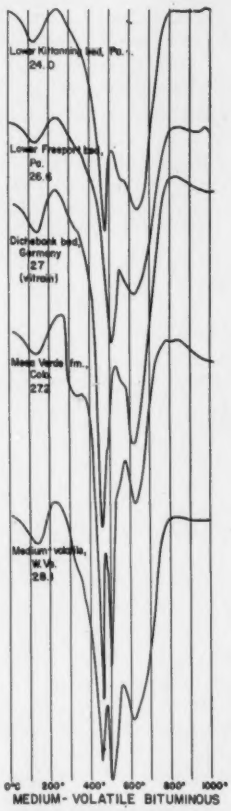


FIG. 3. (*Left*) Low-volatile bituminous.
FIG. 4. (*Right*) Medium-volatile bituminous.

highest V.M. percent shown for semi-anthracite is 11.9 percent (Fig. 2), and the lowest for low-volatile bituminous is 15.9 percent (Fig. 3). The exact V.M. percent for the appearance of the 500° C. peak is, therefore, not evident, but it is possible that it may coincide with the A.S.T.M. boundary value of 14 percent.

The general features of the curves continue their trend towards larger peak sizes and lower temperatures. The low-temperature exothermic peak is between 230–280° C, and the higher-temperature volatile loss peak, equivalent to the single anthracite type peak, lies between 610–635° C. The range for the new lower-temperature volatile loss peak is 490–510° C. The increase in size of the entire curve and the decrease in temperature of peaks are associated with the increase in V.M. In general, as rank decreases, curves become larger and peak temperatures lower.

All low-volatile and most medium-volatile curves show two endothermic peaks for volatile loss. The lower-temperature peak is referred to as primary volatile loss or devolatilization, and the higher temperature peak as secondary loss.

The curves do not indicate any "break" in rank at the boundary with semi-anthracite. It would appear that the transition from semi-anthracite is continuous and that effects related to plasticity cause the appearance of the 500° C peak.

With decreasing rank of low- and medium-volatile coals the 500° C peak increases in amplitude, reaching a maximum value in the medium-volatile rank.

Medium-Volatile Bituminous.—The curves for medium-volatile coals (Fig. 4) show continuing enlargement of the 500° C peak and general growth of the entire curve. The rank boundary between low-volatile and medium-volatile coals is an arbitrary V.M. value, as the curves indicate a continuous sequence. No samples analyzed at 22 percent V.M. could be obtained, but interpolation of the curves indicates that the A.S.T.M. rank boundary at 22 percent V.M. should show a curve with about equal primary and secondary volatile loss peaks. The medium-volatile curves, therefore, have a primary volatile loss peak greater in size than the secondary, and the reverse is true for low-volatile curves. Thermal curves for these two ranks comprise a single type of plastic coals and are referred to as the low-volatile type.

When the rank has decreased to about 27 percent V.M., the curves show a transition to curves of high-volatile coals. This is indicated by the development of a second peak (27.2 and 28.1 percent V.M., Fig. 4) at a lower temperature (455–465° C). The dual nature of the 500° C peak is evidently an incipient development of the strong exothermic reaction which is characteristic of curves for high-volatile bituminous coals at about 400° C. The transition zone from about 27–29 percent V.M. constitutes the range of the best average for coking blends and natural coking coals.

Temperatures for reactions are slightly lower than for low-volatile coals. The low-temperature exothermic reaction occurs between 225–265° C, the primary volatile loss peak between 480–510° C, and the peak temperature for secondary volatile loss is about the same as in low-volatile curves (615–635° C).

High-Volatile Bituminous.—The high-volatile A bituminous curves (Fig. 5) have the greatest amplitude of all ranks of coal and represent the maximum values for curve size. All high-volatile curves are characterized by the pronounced exothermic reaction at about 400° C. This reaction is evidently an

enlargement of the small effect noted in the transition zone of medium-volatile curves. However, in high-volatile curves the split of the 500° C peak is pronounced and complete. The exothermic effect is superimposed upon the 500° C endothermic peak, causing thermal balance. When the effect of the

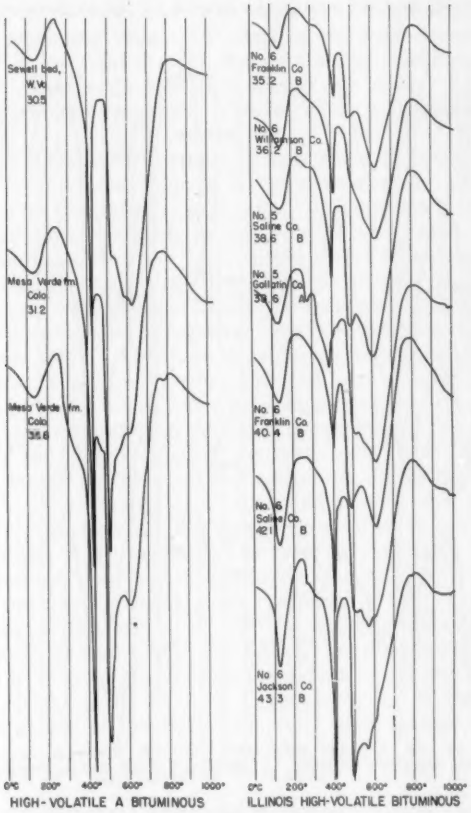

Sewell bed, W.Va. 30.5
Mesa Verde fm. Colo. 31.2
Mesa Verde fm. Colo. 38.6
0°C 200° 400° 600° 800° 1000°
HIGH-VOLATILE A BITUMINOUS
No. 6 Franklin Co. 35.2 B
No. 6 Williamson Co. 36.2 B
No. 5 Saline Co. 38.5 B
No. 5 Gallatin Co. 39.6 A
No. 6 Franklin Co. 40.4 B
No. 6 Saline Co. 42.1 B
No. 6 Jackson Co. 43.3 B
0°C 200° 400° 600° 800° 1000°
ILLINOIS HIGH-VOLATILE BITUMINOUS

FIG. 5. (*Left*) High-volatile A bituminous.
FIG. 6. (*Right*) Illinois high-volatile bituminous.

exothermic reaction is released, the curve then returns to its position on the endothermic effect. This produces a curve with peaks at about 400° and 500° C, the 400° C peak being formed as the exothermic reaction interrupts the continuing endothermic reaction; the 500° C peak forms when the curve is free to return to its position on the endothermic curve.

With regard to primary and secondary volatile loss, the two peaks at about 400° and 500° C constitute the range for primary volatile loss (the range of the 500° C peak in low- and medium-volatile coals), and the 620° C peak represents the secondary. Peak temperatures are slightly lower than for medium-volatile coals. The low-temperature exothermic reaction occurs between 240–250° C, and the two peaks caused by the high-volatile exothermic

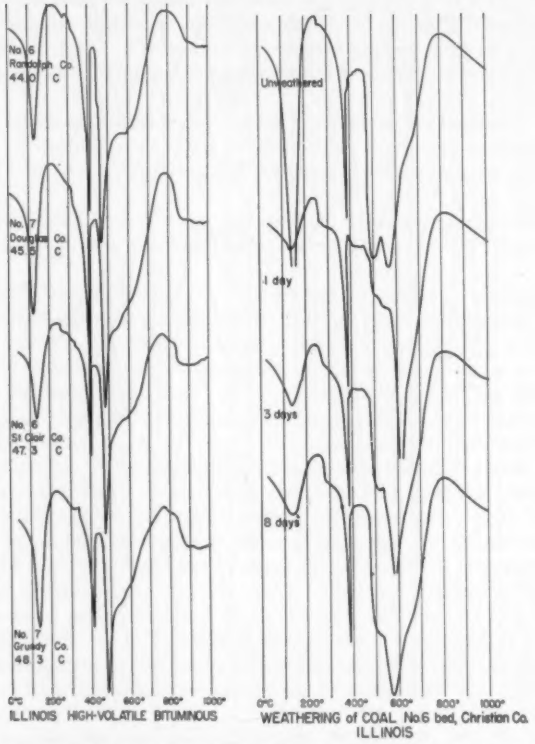


FIG. 7. (*Left*) Illinois high-volatile bituminous (continued from Fig. 6).
FIG. 8. (*Right*) Weathering of Coal No. 6 bed, Christian Co., Illinois.

reaction occur between 425–435° C and 510–520° C. The secondary volatile loss peak occurs between 610–625° C.

Comparison of the lowest rank medium-volatile curve (28.1 percent V.M., Fig. 4) and the highest rank high-volatile curve (30.5 percent V.M., Fig. 5) suggests that the A.S.T.M. rank boundary at 31 percent V.M. is probably slightly high. Interpolation from the curves indicates that the true high-

volatile type may start at about 29–30 percent V.M. The evidence from the curves suggests that the transition between the two ranks is probably gradual. The development of the high-volatile exothermic reaction begins in the transition zone of the medium-volatile rank, and it is to be expected that the reaction gradually increases in intensity until the true high-volatile type is reached.

The thermal curves change as a new physical or chemical property slowly develops with decreasing rank. The start of plasticity causes the 500° C peak to appear in low-volatile coals; the start of the exothermic reaction of unknown cause results in the split high-volatile type curve. As the rank decreases, the high-volatile type curve is succeeded by a subbituminous type as plasticity is lost.

Illinois High-Volatile Bituminous.—The increase in curve size with decreasing rank reaches a maximum in high-volatile A coals. However, coals from different areas give different peak sizes. The problem is further complicated by the fact that weathering in high-volatile coals causes large thermal effects. Weathering effects will be discussed in the following section.

To compare coals from a single area, weathering must be considered and coals must be collected at approximately the same time. Illinois coals used were from a previous study (10) and had been collected more than three years before the present study. They have therefore approximately reached equilibrium with regard to weathering conditions and may be compared with each other; they should not be compared with coals from other areas or with coals from the same area with different lengths of weathering.

The Illinois hv (high-volatile) A coals all come from Gallatin County. Their thermal curves will be found similar to those of the higher rank hv B coals because of their weathered condition and amount of V.M. These hv A coals are therefore included in the discussion of the hv B coals. Curves for Illinois coals (Figs. 6, 7) are shown in order of decreasing V.M. content. They include one hv A coal and representative hv B and hv C coals.

The higher rank hv B coals (35.2–40.4 percent V.M., Fig. 6) are characterized by a strong development of the 620° C secondary devolatilization peak and smaller peaks at about 400° and 500° C for the primary range. The smaller 400–500° C peaks are associated with lower V.M. content, and as the V.M. content increases, the 400–500° C peaks increase in size. There is a slight increase in size of the water-loss peak. This type of curve includes certain Illinois coals which are used in coking blends.

With increase in V.M. content of hv B coals (42.1–43.3 percent, Fig. 6), the 400–500° C peaks increase further in size, water peaks become larger, and the 620° C peak decreases in size and in temperature to about 575° C. This type of curve is representative of the poorer coking hv B coals and is transitional to the hv C coals.

The hv C coals (Fig. 7) show further reduction to final elimination of the peak between 575–620° C, there being no apparent peak development at about 45.5 percent V.M. The 400–500° C peaks are now generally large in size, and the 500° C peak has dropped in temperature to about 480° C. There is, therefore, a smaller temperature interval between the 400–500° C peaks than in hv B coals. Water peaks reach maximum size in hv C coals.

Thus there is a continuous sequence in relative peak sizes and temperatures in Illinois coals with decreasing rank. It is possible by means of thermal analysis to distinguish between hv B and hv C coals, and, with suitable calibration, determination of the approximate V.M. content may be possible. The curves also indicate which coals show the best coking properties as correlated with Illinois State Geological Survey coking coal studies.

The principal features of Illinois curves are listed in tabular form to show the trend in peak sizes.

Rank	Percent V.M.	Primary-secondary volatile loss peaks	500°–400° C. peaks	500°–400° C. temp. interval	Water peak
typical hv B	35.2–40.4	600° > 400°–500°	500° > 400°		small
transitional hv B	42.1	575° > 400°–500°	500° = 400°	90°	moderate
transitional hv B	43.3	575° < 400°–500°	500° = 400°	90°	moderate
transitional hv C	44.0	575° ≪ 400°–500°	500° > 400°	60°–70°	large
typical hv C	45.5–48.8	no 600° or 575° peak	500° > 400°	60°–70°	large

WEATHERING OF COAL

Weathering effects increase with decreasing rank and reach a maximum in the plastic hv C coals. Coals higher in rank than high-volatile weather more slowly and only small changes are observed in the thermal curve.

For purposes of this discussion, No. 6 bed hv C coal (48.2 percent V.M.) from Christian Co., Illinois, was selected. The coal was placed under water at the time of sampling and first exposed to the air at the start of the thermal curve series. The coal was powdered to pass 60 mesh and curves were made on 15 successive days. Exposure to air accelerated weathering of the particles. Curves are shown in Figures 8 and 9; only curves showing an obvious change are presented.

Unweathered coal shows a thermal curve unlike any in the Illinois set in Figures 6 and 7. Such difference shows clearly the impossibility of making valid comparisons between curves unless the possibility of disturbing weathering influences is eliminated. The large water loss peak is caused by the moist fresh sample. Once the sample has dried, water peaks remain constant.

If attention is focused on the two endothermic peaks at 575° and 500° C in the unweathered coal, the trend of the changes due to weathering can be followed. The peak at about 390° C remains fairly constant and need not be considered.

At 1 day a large change has occurred. The 575° C peak has increased greatly in size and is at a higher temperature (615° C). The 500° C peak has decreased greatly in size.

From 3 to 8 days a trend in the shifting peaks can be observed. The 575° C peak decreases in size and drops in temperature, and the 500° C peak increases in size. At 9 days the trend continues, but the 500° C peak is now at 490° C. At 10 days the 575° C peak is practically eliminated, and the

500° C peak has reached maximum development. The change has now reached the stage of the hv C curves of Figure 7.

At 11 days another change takes place in the form of a small development of a 600° C peak. The remainder of the curve is identical with 10 days. The 600° C peak slowly increases in size by 14 days.

The changes beyond 14 days were not studied and therefore the end stage in weathering was not observed. It is to be expected that change would con-

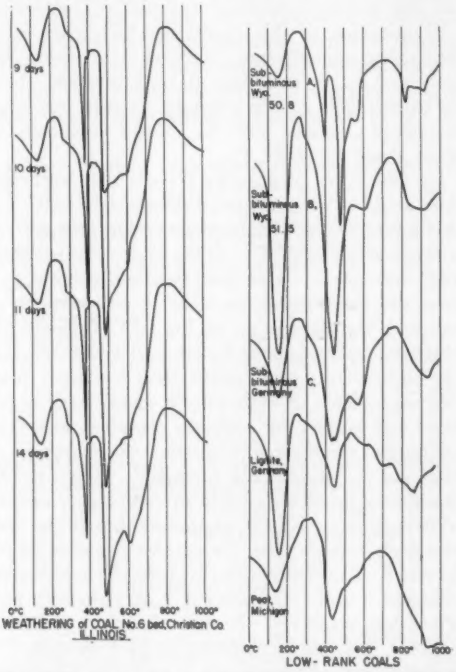


FIG. 9. (*Left*) Weathering of Coal No. 6 bed, Christian Co., Illinois (continued from Fig. 8).

FIG. 10. (*Right*) Low-rank coals.

tinue until the coal has lost its plasticity. At that point the curve would be similar to a subbituminous type thermal curve.

In the weathering of coal, there is a progressive change in the thermal curve except during the first day, when a sudden change takes place. Unfortunately the true significance of the volatile loss changes is not known and no data were available concerning changes in composition, physical properties, and structure. Until such data are available, the fundamental causes for the changes remain unknown.

Interpretation of data from thermal curves, fluidity and swelling measurements, thermal decomposition, and other physical and chemical tests should be made with caution as results may not be significant because of different weathering rates.

LOW-RANK COALS

In coals lower in rank than high-volatile, plasticity decreases and eventually disappears. This is indicated (Fig. 10) by a change in curve type; i.e., the disappearance of the sharp exothermic reaction characteristic of the high-volatile coals. Elimination of the exothermic balancing effect permits the primary devolatilization endothermic reaction to develop freely. Loss of volatiles is represented by an endothermic peak between about 435–445° C, which evidently corresponds to the primary volatile loss peak of low- and medium-volatile curves at about 500° C. Small secondary loss peaks may be present between about 550–600° C.

Thus there are two types of nonplastic coals: the anthracite type with a single volatile loss peak at temperatures between 630–680° C, and what can be called the subbituminous type with a predominant primary volatile loss peak at temperatures between 435–445° C. If the high-volatile exothermic reaction had not occurred, decrease in rank from medium-volatile to subbituminous coals would show the progressive decrease in temperature of the primary volatile loss peak and the elimination of the 620° C secondary volatile loss peak. The subbituminous type curve includes all low-rank coals that are nonplastic as well as high-volatile coals which have lost their plasticity, and the nonplastic banded ingredients of high-volatile coals, such as fusain.

As high-volatile and lower rank coals are classified with respect to heating values, agglutinating and weathering properties, the change in curve type does not correlate with any A.S.T.M. rank boundary. High-volatile C and subbituminous A coals may be either plastic or nonplastic, and each may show curves of two different types. It is suggested that if the thermal curve of fresh coal shows a high-volatile type, the coal be classified in the high-volatile bituminous rank, and if of subbituminous type, in the subbituminous rank.

The subbituminous A coal (50.8 percent V.M.) shows a high-volatile type curve and therefore has plasticity. Because it is not a fresh coal, but still retains some plasticity, it would be classified as hv C in rank.

On the other hand, the subbituminous B coal (51.5 percent V.M.) is not plastic and the characteristic 440° C peak is developed. It is interesting to note the large water content of this coal in spite of its being air-dried for over three years. In general, the low-rank coals have highly variable water peaks.

Curves for subbituminous C, lignite, and peat are of the subbituminous type. The small size of the 450° C peak is caused by excessive shrinkage during firing with corresponding loss in thermal intensity. The volume loss evidently causes the highly erratic high-temperature portion of the curve.

RANK BOUNDARIES

There are five distinct types of thermal curves which form because of the structural, physical, or chemical changes which accompany increasing rank.

Thus there are four natural rank boundaries where thermal types change: (1) between meta-anthracite and anthracite; (2) between semi-anthracite and low-volatile bituminous; (3) between medium-volatile bituminous and high-volatile bituminous; (4) between high-volatile bituminous and subbituminous. The boundaries between anthracite and semi-anthracite and low-volatile and medium-volatile bituminous are gradual and show no change in curve type. Only arbitrary boundaries can be placed at these points.

The thermal curves indicate that coalification is a continuous sequence with increase in rank to anthracite coals. At this point a structural "break" occurs caused evidently by forces not accompanying "normal" coalification processes. The three natural rank boundaries, excluding anthracite-meta-anthracite, do not represent "steps" (11) but only gradually changing properties of the coal, and it is obvious that the gradational boundaries cannot be considered as "steps."

The relationships between rank, thermal type, and plasticity may be tabulated as follows:

Rank	Thermal curve type	Plasticity
Meta-anthracite	Meta-anthracite	Nonplastic
Anthracite	Anthracite	"
Semi-anthracite	"	"
Low-volatile bituminous	Low-volatile	Plastic
Medium-volatile bituminous	"	"
High-volatile bituminous	High-volatile	"
Subbituminous	Subbituminous	Nonplastic
Lignite	"	"
Peat	"	"

TABLE 1

PEAK AND TEMPERATURE RANGES IN DEGREES CENTIGRADE FOR THE RANKS OF COAL

Rank	Water	Low-temperature exothermic	High-volatile exothermic	Endothermic volatile loss	
Meta-anthracite	none	575–610			725–735
Anthracite	125–140	420–470			660–680
Semi-anthracite	130–140	250–350			630–670
Low-volatile bitum.	130–150	230–280		490–510	610–635
Medium-volatile bitum.	125–140	225–265	(455–465)*	480–510	615–635
High-volatile A bitum.	130–135	240–250	425–435	510–520	610–625
Illinois hv B	125–145	220–240	385–420	490–515	570–620
Illinois hv C	120–140	220–240	405–410	465–480	none–590
Subbituminous A**	150	250	400	475	560
Subbituminous B	150	270		445	600
Subbituminous C	150	260		440	570
Lignite	150	250		445	?
Peat	140	320		435	

* Transition zone.
** High-volatile type thermal curve.

SUMMARY

Differential thermal analysis of coal, using the hydrocarbon and other gases lost during heating to exclude air, shows a series of thermal curve types which

depend upon the plasticity of the coal, the organic chemical reactions, and the structural changes which take place when coal is heated in the absence of air.

Five distinct thermal curve types can be recognized, the curve type changing when a structural, physical, or chemical change takes place with change in rank. The classification of curve types is based principally on the number and temperature of endothermic devolatilization thermal peaks.

The meta-anthracite type, which includes only the meta-anthracite rank, is characterized by a single volatile loss peak between 725–735° C. The anthracite type, which includes the anthracite and semi-anthracite ranks, shows a single volatile loss peak between 630–680° C. The presence of only a single volatile loss peak for the three highest ranks of coal is correlated with lack of plasticity.

The low-volatile type curve, which includes the low- and medium-volatile bituminous ranks, is characterized by two volatile loss peaks. The lower temperature peak at about 500° C is referred to as the primary volatilization peak, and the higher temperature peak at about 620° C as the secondary. The presence of two principal volatilization peaks is indicative of plasticity.

The high-volatile type curve, which includes only the high-volatile bituminous rank, may be considered as typical of a second group of plastic coals. The curve is characterized by the sharp exothermic reaction superimposed upon the endothermic primary volatilization peak, resulting in the formation of two peaks. Differences can be recognized between hv A, B, and C coals based on the amplitude of the endothermic peaks. The curves also give a general indication as to the coking ability of these coals.

The subbituminous type curve, which includes the subbituminous, lignite, and peat ranks, is characterized by a large primary volatilization peak at about 450° C. The secondary volatilization peak may be absent or show weak effects between 550–600° C. This type of curve represents the nonplastic low-rank coals.

The thermal curves demonstrate the changes which occur during weathering of high-volatile coals. The greatest change takes place within the first 24 hours and subsequent changes are gradual. The end stage is reached with loss of plasticity, and a subbituminous type curve results.

The five types of curves show four natural rank boundaries where thermal types change. Between meta-anthracite and anthracite a structural "break" in the coalification sequence occurs. The boundary between semi-anthracite and low-volatile bituminous designates a plasticity threshold, and the boundary between medium-volatile and high-volatile bituminous separates bituminous coals with different plastic properties. The boundary between high-volatile bituminous and subbituminous demarks another plasticity change. The boundaries between anthracite and semi-anthracite and low-volatile and medium-volatile bituminous are gradual and show no change in curve type. These boundaries are, therefore, arbitrary.

The natural rank boundaries generally agree with the A.S.T.M. rank classification. However, it is believed that the boundary between medium and high-volatile bituminous ranks should be placed at about 29–30 percent V.M.

Distinctions between hv C and subbituminous A ranks may be based on curve type.

The curves indicate that coalification is a continuous sequence with increase in rank to anthracite. At the meta-anthracite boundary a structural "break" occurs. The remaining three natural rank boundaries cannot be considered as "steps," but are caused by the gradual changing properties of coal with increase in rank. The gradational boundaries obviously cannot be considered as "steps."

ILLINOIS STATE GEOLOGICAL SURVEY,
URBANA, ILL.,
August 1, 1953.

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PYRITIC MINERALIZATION IN THE GOUDREAU AREA OF ALGOMA, ONTARIO

GEORGE VIBERT DOUGLAS

ABSTRACT

The pyritic ores of this area are believed to be of hydrothermal origin. They represent the mineralization by hydrothermal solutions of the sheared and brecciated incompetent layer lying between the competent acid volcanic footwall and the basic volcanic hangingwall. Evidence for these relationships is to be found in the origin of the ottrelite schist of the footwall and in the shearing and veining of the hangingwall. These events preceded the emplacement of the Algoma granite from which the mineralizing solutions are believed to come.

INTRODUCTION

THE Goudreau area is approximately 176 miles from Sault-Ste-Marie on the line of the Algoma Central Railway. The pyritic deposits dealt with in this paper lie to the east of the railway and can be conveniently reached from the Goudreau-Lochalsh road (Fig. 1).

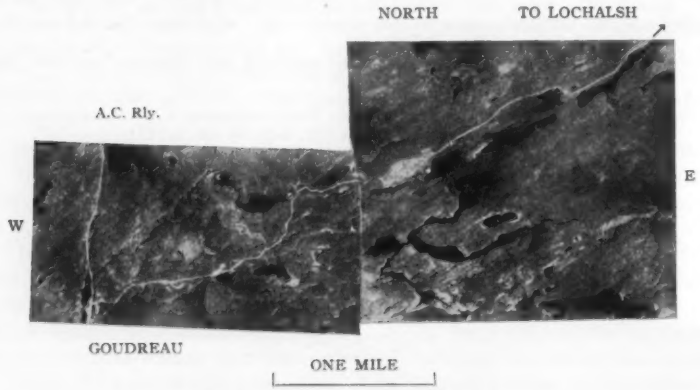

FIG. 1. Air photograph of area.

There have been numerous workers in the field from the days of Sir William Logan and Dr. Robert Bell to the present. It is possible that there are at least forty or fifty references in the literature to these deposits. Michel Goudreau was probably the first prospector in the area about the time of the building of the Canadian Pacific Railway. In 1897 Benjamin Boyer discov-

ered the hematite at what is now the Helen Mine. There was considerable mining activity in the area during the first World War. Between World War I and II interest in gold mining superceded the mining of pyrite. In recent years the area has been dormant.

The purpose of this paper is to discuss the genesis of the pyritic mineralization. The writer is grateful to the management of the Algoma Ore Properties of Sault-Ste-Marie for whom the work was done for permission to present these results.

GENERAL GEOLOGY AND MINERALIZATION

The Goudreau area is on the Canadian Shield. The hills are low and show an accordance of level. The grain of the country is east by north and west by south. In general the grain reflects the geology.

The oldest rocks in the area are acid volcanics that include flows and tuffs, of Keewatin age. The rest of the geological history can be put in columnar form and also is shown graphically in Figure 2.

Latest event	Mineralization	—Introduction of sulfides and carbonates by hydrothermal solutions.
	Igneous activity	—Granites, granodiorite, etc.
	Mountain building	—Folding, brecciation and shearing.
	Sedimentation	—Doré Series.
	Basic Volcanics	—Flows and tuffs.
	Sedimentation	—Banded silica and other sediments
	Acid volcanics	—Flows and tuffs.
Oldest	An earlier granite as seen in the xenoliths found at the Bear Pit (Fig. 4).	

It will thus be seen that these two periods of vulcanism were each followed by erosian and sedimentation; one widespread period of mountain building followed by the customary igneous intrusions and later mineralization.

The facts and observations that support these views are as follows:

1. The mineralization occupies the sheared and brecciated incompetent layer between the acid footwall and basic hanging wall (Figs. 3, 5).
2. The acid footwall has been sheared and in some of its members ottrelite has been developed.
3. The basic hanging wall has been cut by veins of carbonates that carry sulfides (Fig. 6).
4. The carbonates close to the incompetent layer are iron-rich, those at a distance which are found carbonatizing the hanging and footwall rocks are more ankeritic.

These facts will now be considered in more detail.

1. There is no doubt in stating that the volcanics of the Goudreau area have been folded. Nor is there much doubt that the pyritic and sideritic mineral deposits occur between the predominantly acid footwall and the basic hangingwall. In Keewatin time, vulcanism occurred in this part of the Shield,

At first, predominantly acid volcanics and tuffs were poured out over the area. When these rocks cooled, erosian took place and the product was deposited in the low portions of the surface. The evidence for this statement is the fine siliceous siltstone that can be seen at the west end of the Helen Mine, and the banded quartz or silica, a sandstone in the Goudreau area. This interpretation states that the banded sandstone is derived by erosion from these acid volcanics and associated tuffs. In his presidential address to the Geological Society of America the late Professor E. L. Bruce (2)[1] stated: "It is more likely that only the banded silica rock is the true iron formation and that the carbonate and pyrite were formed by metasomatic reactions, not contem-

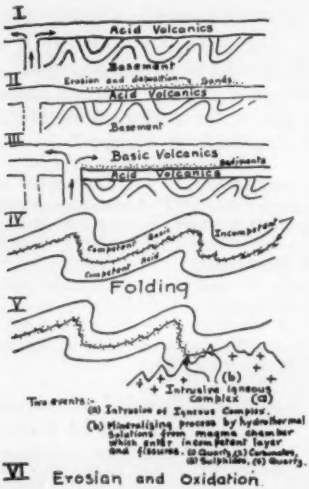

FIG. 2. Diagram of geological events. The Dore Series was laid down on the Basic Volcanics after event III and before IV.

poraneously with the Keewatin rocks, but long after they had been tilted into their present attitude."

Vulcanism again broke out and the basic flows and related tuffs covered most of the area. Some of these flows are ellipsoidal, which supports the idea of the presence of surface waters. Then came another period of erosian and sedimentation, which resulted in the Doré Series. It is possibly significant that the Doré sediments are darker in color than the banded sandstone for these were derived from light colored acid rocks, whereas the Doré came predominantly from darker basic rocks.

The next event was mountain-building resulting in folding and some thrusting. When anything is folded there is differential movement between

[1] Numbers in parentheses refer to References at end of paper.

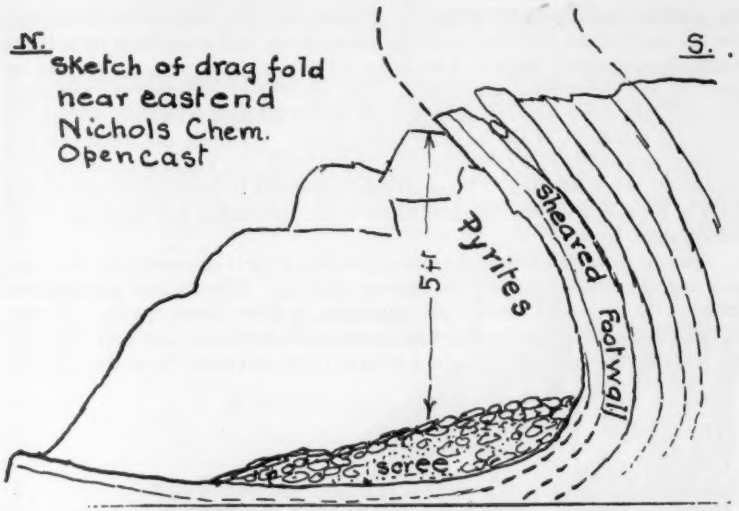

FIG. 3. Nichol's chemical pit. Sketch showing fold in footwall with post-folding mineralization.

FIG. 4. Bear Pit. Xenoliths of granodiorite (white fragments near handle of hammer) in Basic Lava.

the particles making up the material. If there are three layers that are folded, the top and bottom layer being strong (competent) and the middle layer being weak (incompetent) there is a tendency for the incompetent layer to take up most of the differential movement. The amount of differential movement that will take place when these layers are folded through π radians will be $\frac{1}{2}\pi$ times the thickness of the beds on either flank. As there are hundreds and possibly thousands of feet involved in the folding, the amount of differential movement is considerable. Some of this movement is taken up in the minute folds in the fabric of the rock and some in the brecciation and shearing in the incompetent layer.

Igneous intrusion followed mountain-building as is so commonly the case, and mineralization followed the igneous activity. Granite and granodiorite invaded the area and intrude the volcanics and the Doré Series. Hydrothermal solutions, first charged with quartz and carbonates and later with pyrite and quartz emanated from the intrusive rocks and passed into the channel-

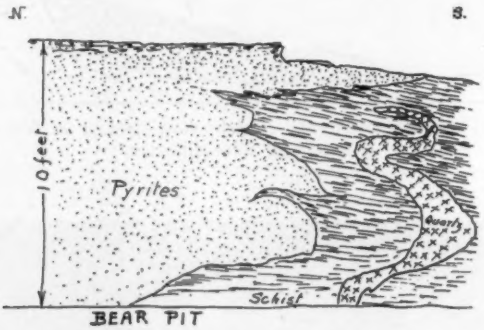

FIG. 5. Bear Pit. Sketch showing pyrites replacing schist.

ways of the brecciated and sheared incompetent layer. Minor paths for these solutions were in shear planes within the volcanics, tuffs and sediments and fissures, which can be seen at various places cutting the volcanics (Fig. 6). These occurrences constitute strong evidence against a sedimentary origin.

2. Collins and Quirke (3) mention the occurrence of ottrelite in the footwall rocks. These authors, after proving that the mineral is ottrelite, one of the chloritoid group of minerals, suggest that this appearance of the mineral represents a new habitat.

Harker (4) discussing the work of Gosselet in the Belgian Ardennes says:

> . . . more striking is the "metamorphisme locale," in which the relation of cause and effect is very clearly displayed. The well-known ottrelite-schists have this origin. They are found in the southern borders of the Cambrian massifs of Stavelot and Serpont, where repeated overthrusts have piled up the Cambrian phyllites and intercalated among them wedges of Devonian Strata. Abundant flakes of ottrelite have been developed in both formations, but only in the near vicinity of the overthrusts. The most interesting localized effects are seen in what

Gosselet styles "metamorphisme par flexion," which stands in close relation to anticlinal or synclinal folds.

This reference removes the peculiarity of the ottrelite occurrence in the Goudreau Area. Its presence has a definite significance. Harker places this discussion of the ottrelite schist in the category of metamorphism brought about by the generation of heat within a rock, which results from the mechanical forces developed by folding or thrusting. The occurrence of ottrelite in the footwall gives some idea of the heat energy developed when those rocks were folded, for there are millions of tons of it in the area. This heat energy

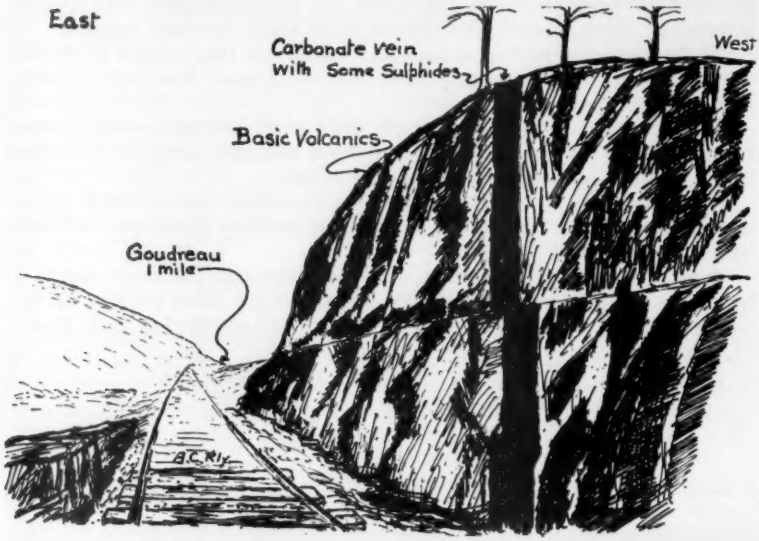

FIG. 6. Sketch showing carbonate vein cutting Basic Lava about one mile north of Goudreau on the Algoma Central Railway.

may also explain some of the facts that those favouring an igneous origin have observed.

3. The carbonate veins and carbonates in the hangingwall and footwall constitute the strongest argument for an epigenetic hypogene origin. Those who advocate a sedimentary origin for the ore have to supply a mechanism whereby the carbonates and sulfides are carried upwards into the basic volcanics and downwards into the acid rocks below the sedimentary ore-bed. Furthermore, it is necessary to time this movement after the basic volcanics were poured out over the ore, which by that theory had been previously deposited.

4. There is further strong evidence in the character of the carbonates.

In general, the siderite is found close to the incompetent layer and ankerite is the common carbonate found in the wall rocks, and in the veins that cut these rocks. This is a broad statement—it does not mean that no siderite is found away from the incompetent layer nor that all the carbonates of this layer are siderite. The statement means that predominantly the siderite is found in the ore zone. This distribution of the carbonates is similar to that in the Fen district of Norway where N. L. Bowen showed that there was a radial distribution with siderite at the centre, dolomite further out, and calcite on the outside.

The present writer visited Norway in 1928 and at that time the Norwegian geologists, Brøgger and Goldschmidt held that these carbonates were igneous injections. Since Bowen's visit little is heard about carbonate magmas.

If the pyrite had been of sedimentary origin, and thus formed before the extrusion of the basic lavas and before folding, it would have been severely brecciated and slickensided.

In one of the Goudreau deposits the shape of the ore-body viewed in section is decidedly puckered but the pyrite shows no slickensiding, which means that it was introduced after the rocks were folded.

The pyritic deposits of Huelva, such as Rio Tinto, are faulted in places and adjacent to these faults the pyrites exhibits excellent slickensided surfaces with a high degree of polish.

It is clear from this discussion of the evidence that the writer favors a hydrothermal origin for these deposits, which does not differ widely from his interpretation of the views held by E. S. Moore (5) and the late E. L. Bruce (1, 2).

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HALIFAX, NOVA SCOTIA,
Oct. 5, 1953.

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ANTHOPHYLLITE–VERMICULITE DEPOSIT OF HAFAFIT, EASTERN DESERT, EGYPT

M. S. AMIN AND M. S. AFIA

ABSTRACT

The gneisses of Hafafit enclose small serpentine masses which are traversed by feldspar veins. On either sides of these veins, vermiculite and anthophyllite are formed in workable quantities.

INTRODUCTION

THE anthophyllite-vermiculite deposit examined occurs in Wadi Hafafit in the Central Eastern Desert of Egypt (Fig. 1). The deposit was briefly examined by the first author in 1947, but in 1952 it was possible to arrange for the detailed examination of the deposit by both authors.

The deposit is exploited at present by the Upper Egypt Mining Co. S.A.E., who kindly permitted the examination, and placed their chemical analyses at the authors' disposal.

GENERAL GEOLOGY

The deposit occupies the lowlands on either side of Wadi Hafafit, which are surrounded by the massifs of Gebel Migif and G. Um Kharaba, which is an extention of the Hafafit Range. These massifs were previously examined by Hume (3,[1] p. 63) and Ball (1, p. 333) who considered them to be made up of granite and hornblende gneisses of the Precambrian basement complex of Egypt.

Little opportunity was available to study these huge massifs, but the following gneissic rocks were recorded during the general reconnaissance of the area: hornblende gneiss, banded biotite gneiss, granitic gneiss, garnetiferous granulite. Some of these gneissic rocks are strongly foliated or banded and commonly possess preserved sedimentary structures indicating their sedimentary origin.

The area of the anthophyllite and vermiculite deposit is almost surrounded by the gneissic massifs of G. Migif, G. Um Kharaba, and G. Hafafit. These form a remarkable huge dome structure whose northern part is readily seen at the watershed of Wadi Hafafit. Unfortunately, the regional geological structure of the country is not yet mapped, but an attempt was made to plot the regional strike and dip of the dome structure on the topographic map of the area surrounding the deposit. This shows (Fig. 2) the closure of the structure on the northern, eastern, and western sides, but it could be followed southward. The dome structure is generally bisected by Wadi Hafafit which runs for a considerable distance in a NNW–SSE direction.

[1] Numbers in parentheses refer to References at end of paper.

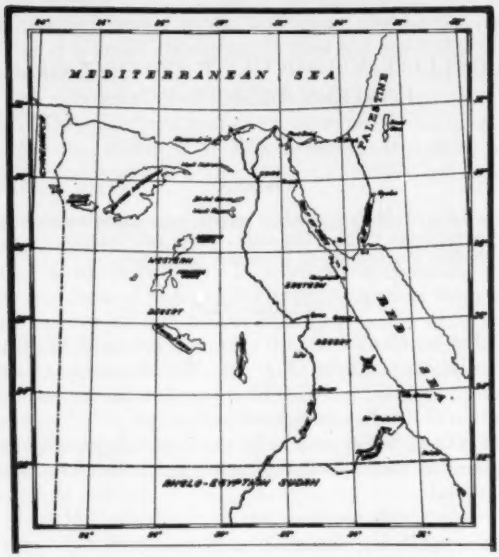

FIG. 1. Map of Egypt showing the locality of the deposit marked by X.

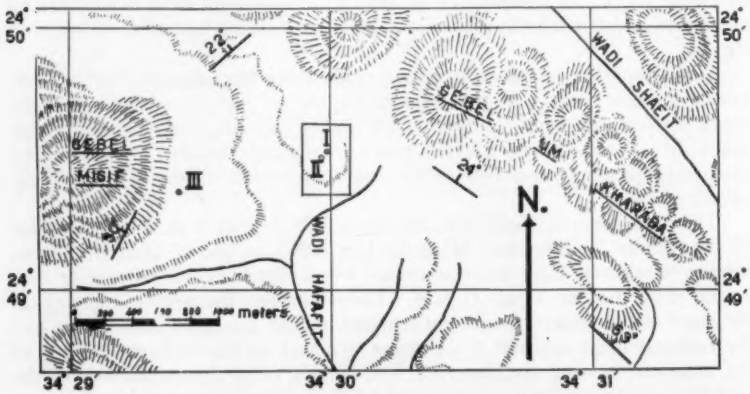

FIG. 2. Map of the country around the deposit. Localities I, II, and III are marked, and the area mapped geologically is in the oblong.

GEOLOGY OF THE DEPOSIT

The lowlands in which the anthophyllite-vermiculite deposit occurs are principally made of granitic gneiss, hornblende gneiss, transformed serpentine masses, and feldspar or pegmatite veins (Fig. 3).

The granitic gneiss is the chief country rock. It is a fine-grained and deeply weathered gneiss, consisting essentially of quartz with undulose extinction, sodic plagioclase, some orthoclase and biotite. The foliation has a general northwesterly trend (Fig. 3), and a gentle dip.

The gneiss encloses numerous flattened lenses of serpentine. In plan, these are irregular or roughly circular (Fig. 3). They are apparently intercalated with the foliated gneiss, with conformable dip and strike. The ser-

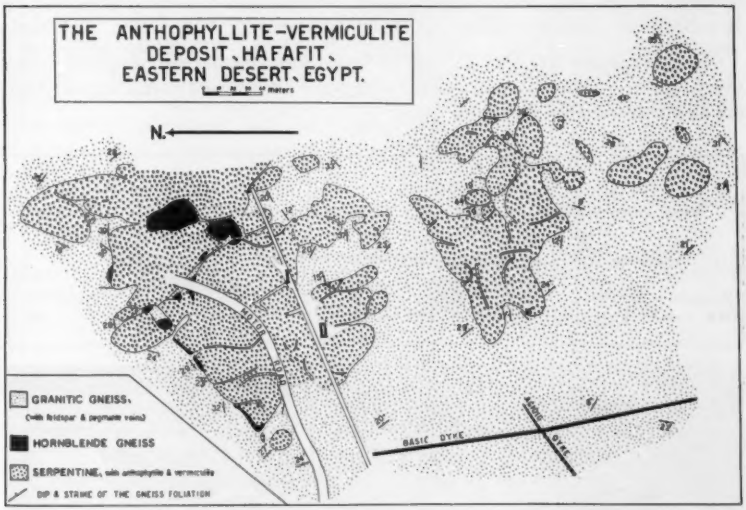

FIG. 3.

pentine rocks are much altered, and are of light or dark brown color, with admixed limonite and carbonate. The least transformed varieties of the serpentine rocks are of dark brown color and consist essentially of serpentine meshes, tremolite, and magnetite with a few relics of olivine.

The hornblende gneiss is a minor formation and occurs as small lenticular masses near the serpentine. It is readily distinguished by its dark green or black color. It is a medium grained rock consisting of hornblende, clinozoisite, saussuritized plagioclase and a little carbonate.

The above formations are commonly traversed by numerous feldspar and pegmatite dikes which weather to the same color as the granitic gneiss and are difficult to distinguish from them in mapping. Within the serpentine and

the hornblende gneiss, the dikes consist of milky feldspar without visible quartz. However, these dikes are quartzose within the granitic gneiss, and typical pegmatite dikes generally associated with the granitic gneiss in the area. The dikes vary greatly in strike and thickness.

The feldspar dikes are of milky white color and sintery appearance. They consist essentially of albitic plagioclase, which is only slightly kaolinized and sericitized, and there may be present some vermiculite, chlorite, corundum or a little garnet. The anthophyllite and vermiculite occurs in close association with the serpentine and feldspar dikes, notably in along the walls of the dikes within the serpentine masses. Several occurrences of the deposit were examined of which the following three examples are described in detail, two of which are from the eastern side of the area and the third is from the western side.

(Figs. 4A & B) EXAMPLE I

In this locality (I, Fig. 2), the feldspar dikes dissect the serpentine into ovoid masses. These, are associated with some hornblende gneiss lenses which are also traversed by the dikes.

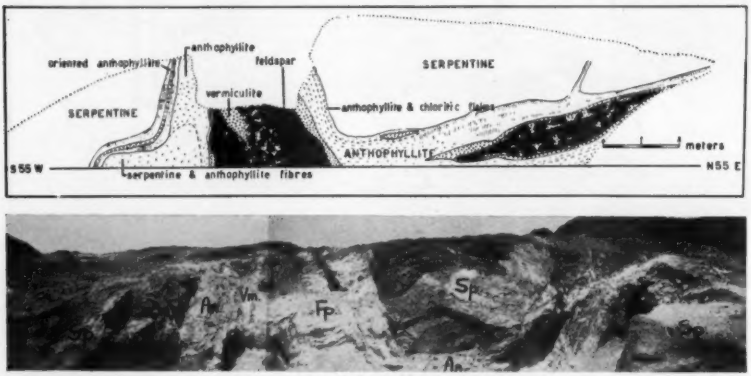

FIG. 4. A cross-section of the anthophyllite-vermiculite of the deposit at locality I showing the distribution of serpentine (Sp), anthophyllite (An, vermiculite (Vm, and feldspar (Fp).

The serpentine is generally of dark brown color. It may be veined by thin magnesite veinlets or exhibits a vesicular appearance due to strong silicification. However, close to the feldspar dikes, the serpentine becomes soft, has a pale green color, and genuinely gives way to anthophyllite on either side of the feldspar dikes.

The relatively fresh serpentine rock of this locality is of dark brown color. It consists essentially of serpentinized olivine grains, chlorite, and tremolite. The original mesh streucture of the olivine is well preserved, being marked by veinings of carbonate and limonite. Numerous chlorite pseudomorphs

are present and they may also enclose a few relics of tremolite. Some magnetite and talc flakes are scattered throughout the rock. The serpentine is therefore considered to be derived from a peridotite parent rock.

The anthophyllite and vermiculite deposit in this locality is extensively worked and provides a good cross-section (Fig. 4A & B), which shows three main zones—serpentine, anthophyllite, and feldspar-vermiculite zones.

The serpentine zone is a much altered rock around and between the anthophyllite zone. It is a friable grayish-brown rock, mixed with yellow-stained carbonate grains and with scattered light brown and greenish flakes of a micaceous mineral. The rock is so friable that it is hardly possible to prepare a thin section out of it.

The anthophyllite zone lies between the serpentine and the feldspar-vermiculite zone. The anthophyllite is a uniform massive rock whose fibers in places may be oriented perpendicular to the serpentine contact. It consists of milky white fibers, with which may be associated a few flakes of brown vermiculite. The anthophyllite zone at the base of the cross-section is a soapy schistose rock, locally known as "Malas." This is essentially made up of anthophyllite, talc flakes, and some vermiculite.

The contact of the serpentine and anthophyllite zones is variable; on the left-hand side of Figure 4A, the two zones merge into each other forming a subsidiary zone of serpentine with anthophyllite fibers. On the right-hand side, the serpentine is in sharp contact against the anthophyllite.

The feldspar-vermiculite zone consists of pockets, narrow seams, or a sheath of vermiculite on either or both sides of the feldspar. On one side of the feldspar dike, the vermiculite contains anthophyllite fibers, but generally the vermiculite is almost free from anthophyllite.

(Fig. 5A & B) EXAMPLE II

The serpentine of this locality is in the form of flattened ovoid bodies, dissected by several feldspar dikes (Fig. 3). The anthophyllite and vermiculite in this locality are mined from a narrow trench along one of the feldspar dikes, which discloses again a serpentine, an anthophyllite, and a feldspar-vermiculite zone.

The serpentine zone consists of a cavernous light brown serpentine cap rock with some relics of dark brown serpentine. The later consists of mesh serpentine, tremolite, and some platy pseudomorphs of chlorite (Fig. 3); some magnetite and spinel are also present. The rock is partly hematitized, and encloses some carbonate grains. The light brown cavernous serpentine cap consists of hematitized mesh serpentine intersected by carbonate veinlets, and contains some talc flakes. The serpentine cap merges gradually into a highly carbonatitized grayish-brown rock made up principally of carbonate and relics of serpentine meshes; some tremolite and talc flakes are present as well.

The anthophyllite zone is in sharp contact with the adjacent serpentine-carbonate rock. The anthophyllite fibers are mostly unoriented but in places form cross-fibers perpendicular to the serpentine contact. The anthophyllite

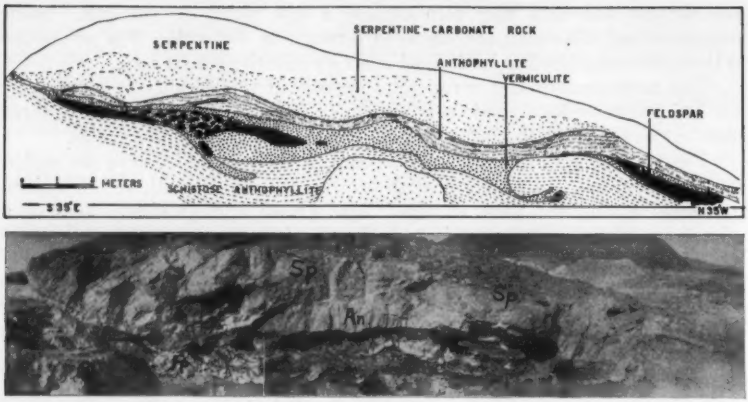

FIG. 5. A cross-section of the anthophyllite-vermiculite of the deposit at locality II, showing the distribution of serpentine (Sp), anthophyllite (An), vermiculite (Vm), and feldspar (Fp).

rock is of milky white color, with a few dark specks of iron oxides. A few flakes of brown vermiculite are scattered in between the anthophyllite fibers. The anthophyllite at the base of the cross-section forms a strongly schistose and compact rock. In places are small kernels of feldspar, around which the anthophyllite is moulded and contorted.

The feldspar-vermiculite zone lies in sharp contact in the anthophyllite. The feldspar and the vermiculite of this zone may form mono-mineralic masses or may be thoroughly mixed together (Fig. 5A). The feldspar rock is of milky white color with a sintery appearance, and consists almost essentially of albitic plagioclase, with a few flakes of chlorite and vermiculite. The plagioclase is slightly kaolinized and sericitized. The vermiculite is flaky and ranges in color from light brown to normal brown. In places it is associated with small seams of olive green chloritic or micaceous mineral.

(Figs. 6A, B, & C) EXAMPLE III

The quarrying of the anthophyllite and vermiculite associated in this locality has exposed the boundaries of the serpentine (Fig. 6A). The feldspar dikes are also exposed in various places around the serpentine mass. The best cross-section is along the main trench, where a complete succession of serpentine, anthophyllite, feldspar-vermiculite and granitic gneiss can be seen (Fig. 6B). This example differs from the previous ones in the occurrence of the feldspar-vermiculite and anthophyllite zones at the contact of the serpentine and the granitic gneiss. The cross-section shows the following zones: 1, serpentine; 2, anthophyllite; and 3, feldspar-vermiculite-actinolite-tremolite.

The serpentine zone is on the right-hand side of the section. The least altered parts of the serpentine zone (off the section) consists of serpentinized olivine and tremolite (Fig. 7), the latter being replaced by talc flakes. The altered serpentine on the right-hand side of the section is strongly veined by magnesite network.

The anthophyllite zone grades outward irregularly from the serpentine-magnesite zone through a wide mass of impure serpentine-anthophyllite rock which consists of relatively coarse fibers embedded in a fine matrix of tufty

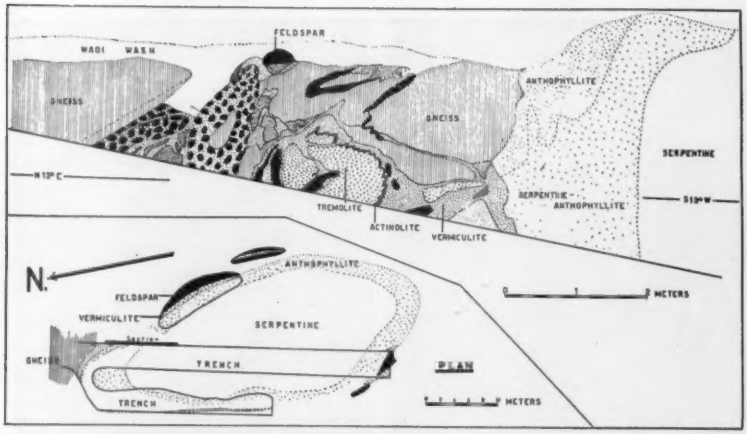

FIG. 6a. Plan of the deposit at locality III.
FIG. 6b. Sketch of cross-section of the anthophyllite-vermiculite deposit along the main trench.

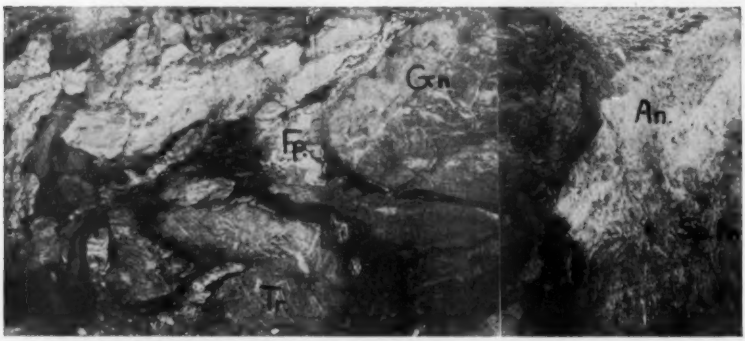

FIG. 6c. Panorama of the same cross-section showing the serpentine-anthophyllite (Sp, Am), anthophyllite (Am), vermiculite (Vm), tremolite (Tr), gneiss (Gn), and feldspar (Fp).

anthophyllite, carbonate and hematitized serpentine (Fig. 8). The white anthophyllite rock is of the normal mass fiber type, with a few specks of vermiculite and stains of hematite.

The feldspar-vermiculite-actinolite-tremolite zone is rather complicated, due principally to its position between the serpentine and anthophyllite zones on one side and the granitic gneiss on the other side. The gneiss forms kernels among the feldspar, vermiculite, tremolite and actinolite masses of the zone. They are also strongly deformed which adds to the complications of the zone.

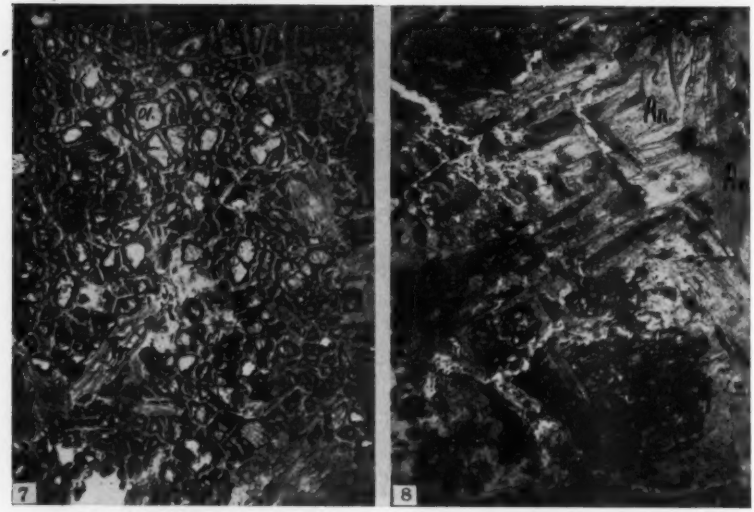

FIG. 7. Photomicrograph of serpentine composed of olivine relics (Ol.) enclosed in a hematitized serpentine matrix, with some tremolite fibres (Tr) partly replaced by talc. × 16.

FIG. 8. Photomicrograph of a serpentine anthophyllite rock made of tufty anthophyllite (An) and hematitized serpentine (Sp). × 16.

The feldspar rock forms detached lenticles or contorted seams. It is generally associated, with various proportions of deep brown vermiculite, which may also form monomineralic pockets of various sizes. In this occurrence, the vermiculite is commonly associated with actinolite and tremolite, which is rather exceptional. The actinolite is of olive-green color and with $Np = 1.630$. The tremolite is of pale green color and with $Np = 1.620$. Both the actinolite and tremolite form irregular shaped masses around the vermiculite or dispersed in it.

PETROGRAPHY OF THE ANTHOPHYLLITE AND VERMICULITE

Anthophyllite.—In these occurrences, the anthophyllite consists of tufty mass fibers that rarely exceed 4 mm in length. The fibers are generally un-

oriented and interlacing, with here and there a tendency for radial arrangement. The character of the anthophyllite makes it suitable only for insulation purposes in cement or plaster. It is of no economic value for spinning purposes. The fibers have positive elongation with straight extinction, and Np = 1.605 (in ordinary light). It is also optically positive.

A typical average chemical analysis of the anthophyllite, as given by the Upper Egypt Mining Co., is:

	%
Ignition loss at 1000 c	5.4
SiO_2	52.3
Al_2O_3	2.1
Fe_2O_3	7.2
CaO	2.7
MgO	28.1
CO_2	2.2

Vermiculite.—The vermiculite is a medium-grained micaceous mineral ranging in color from coffee brown to light brown to bronzy. The mineral exfoliates upon heating due to the release of water. The heated vermiculite is transformed into submetallic lustrous flakes with paler color than the unheated material, and becomes lighter and voluminous. The heated flakes are also rather opaque, possibly due to the oxidation of iron oxide that appears as dense fine dust throughout the flakes.

The optical properties of two varieties of vermiculite are given below:

	Coffee brown	Brown
Nm =	1.619	1.605
2 V(−)	17°–18°	0–5°

The green chloritic or micaceous mineral commonly found in association with vermiculite possesses strong birefringence with 2 V (−) = 0 − 10 (variable). The average chemical composition of the brown vermiculite, as given by the Upper Egypt Mining Co. is:

	%
SiO_2	36.23
$Fe_2O_3 \cdot Al_2O_3$	20.58
CaO	1.54
MgO	23.59
Loss on ignition	18.36

The analyzed sample was dried at 100° C.

DISCUSSION

The study of the previous examples of anthophyllite and vermiculite deposits permits conclusions as to the general relations of these deposits to the associated country rocks.

The deposits are confined to the serpentine masses associated with the gneisses of Wadi Hafafit. These serpentines are cut by feldspar dikes. This occurrence indicates the genetic relation of the anthophyllite and vermiculite deposits to the serpentine and associated gneisses and feldspar dikes. Similar relations are recognized for anthophyllite and vermiculite to altered ultra-

basic rocks in several other localities, such as Shetland Islands and Lizard of the British Isles, Palaboroa in South Africa, and in Rainy Creak of Montana.

In the Shetland Island, C. Phillips (5, p. 643) recorded the formation of anthophyllite followed by talc and actinolite at the contact of serpentine and gneisses. In the Lizard, J. S. Flett and J. B. Hill (2, p. 106) also described the formation of anthophyllite together with talc, tremolite, and chlorite at the contact of serpentine and gneisses. In the famous Palaboroa area of South Africa, C. M. Schwellnus (6) described the formation of a vermiculite deposit at the contact of pyroxenite and syenite rocks. In Rainy Creak of Montana, J. T. Pardee and E. S. Larsen (4, p. 24) described the formation of vermiculite with altered pyroxenite.

It is also noteworthy that the anthophyllite and vermiculite of Hafafit are strictly confined to the walls of the feldspar dikes within the serpentine masses. These dikes extend also into the granitic gneiss where they form pegmatite dikes, but they contain neither anthophyllite nor vermiculite, and only a few specks of biotite. This indicates that the formation of anthophyllite and vermiculite in Hafafit is closely related to the occurrence of feldspar dikes in the serpentine. The formation of these dikes, as well as of their extensions into pegmatites in the granitic gneiss, is not an uncommon phenomena and is undoubtedly related to the process which led to the formation of the gneisses themselves. These dikes be either sweated out fluid produced by selective refusion of the country rock, or they may be late granitic injections.

The confinement of anthophyllite and vermiculite to the contact of feldspar and serpentine and their absence from the pegmatites in the granitic gneiss is significant. This indicates the genetic relation of this deposit to both serpentine and feldspar. The vermiculite is a Mg-Fe-Al-rich mineral, and the anthophyllite is a Mg-Fe-Si-rich mineral. Both require a Mg-Fe source and Al-Si source. These can be provided by the serpentine and feldspar respectively.

The field relations of the anthophyllite and vermiculite deposits and the associated serpentine and feldspar dikes at Hafafit show, as previously pointed out, a fixed relationship of feldspar dikes, paralleled by vermiculite, anthophyllite, serpentine-carbonate rock and serpentine. The vermiculite is mostly confined to the walls of the feldspar dikes forming pockets or narrow seams or sheaths bordering the dikes. In places, the vermiculite is disseminated as well. Away from the dikes, the vermiculite is found only as a few specks in the anthophyllite zone or in the serpentine-carbonate rock on the sides of the feldspar. Unfortunately the vermiculite zone does not lie in direct contact with the serpentine but is separated from it by the anthophyllite zone.

The vermiculite in places is associated with tremolite and actinolite, but only in example III where feldspar dikes inject the contact of the serpentine and the gneisses, instead of being confined within the serpentine mass. The formation of actinolite and tremolite might be attributed to the effect of gneiss.

The contact of the anthophyllite and the vermiculite-feldspar zones is rather sharp, but that of anthophyllite and serpentine is either sharp and marked by the formation of cross-fibers anthophyllite, or is a gradational contact and merges through a serpentine-anthophyllite subzone. The antho-

phyllite is generally made of interlacing mass fibers with a few cross or slip fibers. This general habit indicates the absence of directive tectonic influence during the formation of this zone. The formation of the cross fibers close to the serpentine contact is local and is possibly due to fracture filling. The slip or schistose fibers of the anthophyllite (example III) are tectonic phenomena but they are very localized also.

The formation of the anthophyllite at the contact of serpentine and feldspar is not an uncommon phenomena, as described by Phillips (5) and Flett (2). It is a (Mg, Fe) rich mineral and is possibly derived from ultrabasic parent through the addition of acidic foreign material. Such additions can be provided by the feldspar injections.

The serpentine rock at the contact of the anthophyllite is strongly veined by carbonate veinlets, with a little talc, green mica, and a few vermiculite flakes. This highly altered variety merges gradually into a normal fresh serpentine. The latter is microscopically proved to be derived from a peridotite made up of olivine and calc-magnesian silicate.

The following conclusions may be drawn regarding the origin of these mineral deposits: The serpentine masses included within the granitic gneisses of Hafafit were invaded by feldspar dikes, which extend into the granitic gneiss as pegmatite dikes. The injection of these dikes was apparently accompanied by a strong flow of hydrothermal fluids. These permeated the serpentine masses on either side of the feldspar dikes leading to the formation of vermiculite close to veins, followed by anthophyllite towards the serpentine contacts.

DEPT. OF MINES & GEOLOGICAL SURVEY,
CAIRO, EGYPT.
July 18, 1953

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EXPENSE OF EXPLORATION

H. M. CALLAWAY

ABSTRACT

An attempt is made to express by formula the factors determining permissible cost of exploration, and hypothetical exploration problems are discussed to demonstrate how the formula may be used to clarify thought on prospect analysis.

INTRODUCTION

So intimate is the relation of geology and economics in the mining industry that their divorcement is impossible. The method of approach to the successful ore search program reflects the skillful blending of the two.

The following discussion of the permissible cost of exploration was inspired by the paper of Thomas B. Nolan entitled "The Search for New Mining Districts," [1] and may be considered an economic supplement to this thought-provoking paper.

FACTORS DETERMINING PERMISSIBLE COST

Obviously, there is a limit to the number of dollars that can be economically spent in search of an ore body of a given value. Since the value of the ore body is related to the cost of attempting to find it, every effort should be made to estimate its probable net yield in the event of discovery. The fact that a hidden ore body can not be accurately evaluated does not exempt one from the responsibility of approximating its value; rather, the situation stresses the need for use of all available information concerning the district or area in which exploration is to be undertaken. A diligent search of published material concerning past production figures will often disclose enough pertinent facts to infer a value for such ore bodies that may remain, as yet, undiscovered.

The amount spent in ore search should also be commensurate with the chances of making a discovery. The quantity of chance is more difficult to determine than that of value, it being dependent chiefly on geologic theory. Past exploration activities within a given district may, however, be taken into helpful consideration. The accuracy of the estimation of chance is directly proportional to the amount of geologic knowledge of an area. Thus, in an unexplored district where geologic information is lacking, the odds for discovery are one-in-many thousands. Within an expanding mining field where geologic controls are well known the odds may be in the order of one-in-twelve.

Summarizing, then, the cost of exploration varies directly as the anticipated value of the discovery and as the chance of making the discovery.

[1] Econ. Geol., vol. 45, no. 7, 1950.

Let C = the maximum number of dollars that can be economically spent on a given exploration project.

R = the estimated chance of making a discovery from the wise spending of C dollars. R is expressed as a fraction whose numerator is the number of discoveries and whose denominator is the total number of attempts to make a discovery. Example, if one drill hole in twenty hit ore, then $R = 1/20$.

N = the net dollar value that can reasonably be expected from the discovery.

Then, $C = RN$.

It is understood that numerical values cannot be substituted in the formula, thereby obtaining a mechanical answer. This fact need not seriously impair its usefulness. The formula is capable of classifying exploration projects into their relative positions of worthiness.

That R and N are both factors of equal weight should be recognized. For example, suppose there are two exploration projects under consideration. One is in a partially developed ore field where geologic controls are well known and past production records indicate the ore bodies to be fairly low grade. Past exploration indicates that one hole in twenty may be expected to lead eventually to development of an ore body of such size that it will net $80,000. The second project is in a new area where a recent high-grade strike has been made, but geologic controls are not understood. The chances of making a discovery by randomly prospecting the new area are believed to be 1/500. The operators of the newly discovered strike value their interest at $2,000,000.

In the first case:

$$C = RN = 1/20 \times 80{,}000 = \$4{,}000$$

In the second case:

$$C = RN = 1/500 \times 2{,}000{,}000 = \$4{,}000$$

Then, disregarding the "luck" factor which is entirely undependable, the two exploration projects have equal economic merit.

THE PROGRESSIVE NATURE OF EXPLORATION SPENDING

Exploration is a calculated risk which must be taken if depletion of present reserves is to be arrested. Though it is indeed expensive, the progressive nature of the investment puts it in the realm of sound economics.

An ore body recently discovered is reported to have cost over $2,000,000 in exploration expense. When prospecting was initiated there was very little geologic criteria to indicate ore. Does this mean that an organization who desires to hunt for an ore body under similar circumstances must be willing to gamble $2,000,000 on odds that are tremendously unfavorable?

Suppose the chances of discovery are one-in-a-thousand and the probable net value of the discovery will be $8,000,000.

$$C = RN$$

$$C = 1/1000 \times 8{,}000{,}000 = \$8{,}000$$

It is, then, economically sound to spend $8,000 on an ore-search program. The amount is small and will limit activity to surface reconnaissance and possibly some detail study of favorable small areas.

By the time this first phase of prospecting is completed, there will have accumulated sufficient information to re-evaluate the problem. If R has a lower value, then the program should be concluded. However, if the pattern of occurrence of the ore is indicated by the reconnaissance, the value of R may rise from 1/1,000 to 1/300.

$$C_1 = R_1N - C$$
$$C_1 = 1/300 \times 8{,}000{,}000 - \$8{,}000$$
$$C_1 = \$19{,}000$$

C_1 is now the justifiable expense of further exploration and is adequate to sustain a limited drilling program or a geophysical survey.

Successive phases of re-evaluation follow in like manner. At the end of every phase the chances of discovery are reconsidered and the maximum justifiable expense for further exploration calculated. At any time that the value of R declines the project may be concluded. The end-point of a successful program would be at a phase when R reached a very high value so that the actual cost of "blocking out" ore would be far less than the maximum allowable cost of drilling, C.

The above example illustrates the progressive nature of exploration spending. Though the cumulative cost of exploration may be $2,000,000, the end-point is proved ore of sufficient grade and tonnage to yield a profit. At no time during the execution of the program would there be money spent that would not be economically justified.

The empiric will find argument in that the chance quantity, R, and the value quantity, N, cannot be accurately determined. Though many formulas can be used mechanically, their primary function is to express specific relations between quantities. It is not essential that we know their absolute quantity in order to examine how they vary *relative* to each other. Since the problem of exploration cost is concerned with the *relative* merits of two or more prospects, or with the *relative* value of one prospect at successive stages of exploration, then the use of the formula, $C = RN$, seems adequate for the logical analysis of permissible exploration cost.

EAGLE-PICHER CO.,
MORRISTOWN, TENN.,
Oct. 14, 1953.

DISCUSSION

SOME CONSIDERATIONS REGARDING LIQUID INCLUSIONS AS GEOLOGIC THERMOMETERS

Sir: The above paper (B. J. Skinner, 1953, ECON. GEOL., vol. 48, no. 7, pp. 541–550) gives a number of interesting facts, but does not answer the question as to whether or not all liquid inclusions in minerals lose or gain fluid when there is a pressure difference between the fluids in inclusions and outside the host crystals.

Skinner heated for a long time one specimen of quartz to a temperature above that of filling of the inclusions by the liquid phase and afterwards found the temperature of filling to be raised slightly (14.4% in 3 hours, 15.6% in 20 hours, 15.3% in 110 hours). On the other hand, he heated for a long time another specimen to a temperature below that of filling by the liquid phase and afterwards found the temperature of filling to be lowered about 10 percent. The fact that the external pressure was one atmosphere in the first set of tests, and 600 and 850 atmospheres in the second set may or may not be significant. The experimental fact that lower preheating temperatures give lower filling temperatures and that higher preheating temperatures give higher filling temperatures should be investigated independent of the other variables, external pressure and time. This important matter of leakage of inclusions cannot be considered to be settled by the results of looking at 33 inclusions after treatment of the host to various mixed temperature, pressure, and time treatments.

Skinner raised once again the fact of variability of ratio of gas to liquid as one which indicates that fluid must have entered or left some of the inclusions since crystallization of the host. Studies related to variability of ratio have been carried out since the middle of last century, and the results appear to be consistent.[1] The variability of ratio is due to inconstant ratio of carbon dioxide to water in the inclusions, and quartz with inclusions of this type is most often reported from pegmatites and "high temperature" quartz veins. The inconstant ratio is undoubtedly related to the nearly immiscible relations between water and carbon dioxide when both are in a dense fluid state, and the current hypothesis is that the two were included in the growing quartz as a mixture of partly miscible fluid phases. Identification of considerable amounts of carbon dioxide in inclusions can be made by observing a third fluid phase, which usually necessitates cooling below room temperature.

Three phase fluid inclusions (two liquids plus a gas), apparently primary, are frequently seen in quartz veins of the Canadian Shield. Also, three phase

[1] Smith, F. G., 1953, Historical development of inclusion thermometry: University of Toronto Press.

saline inclusions (a liquid plus a salt crystal plus a gas), apparently secondary, are present in planar groups. The two kinds of inclusions, with very different composition and pressure (the primary type is under a pressure slightly below the critical pressure of one of its components), may be seen separated by distances of the order of 1 micron. The pressure gradients must be of the order of 40 times greater than those in Skinner's experiments and the time of disequilibrium, of the order of 10^{10} times greater.

In this laboratory, inclusions in a large number of minerals have been studied in 1) polished thin sections with oil-immersion methods, 2) polished thick sections in a heated microscope stage, and 3) sized fragments in the decrepitation apparatus. The results that bear on the matter of leakage (in or out) of inclusions indicate that when a crystal is sawed or crushed, some of the inclusions will leak nearby the fresh surfaces, but not all of the inclusions will leak until the internal pressure is very high. (Of course, if the host crystal were able to retain inclusions at any temperature above that of filling, the decrepitation method could not be used.) The smaller the inclusion, the greater is the probability that it is surrounded by the continuous crystal matter of the host, and therefore could not leak until rupture of the host. The huge size of the inclusions studied by Skinner, up to 1.25 mm in length (within the sections of quartz only 1.50 mm thick), can be considered to be mineralogically abnormal. The mean size of fluid inclusions for good decrepitation data is between 1 and 10 microns. These are within fragments about 0.25 mm across, that is, much smaller than major lineage elements and possibly smaller than most lineage elements.

Many of us will continue to regard fluid inclusions as good samples of fluid present during crystallization, and permanently recording its density at that time.

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UNIVERSITY OF TORONTO, CANADA,
December 7, 1953

THE KOLHAN SERIES, AND BANDED HEMATITE QUARTZITE [1]

Sir: May I apologize for asking you to publish this letter as it may appear to continue a not very useful discussion. However, Drs. Spencer and Percival in their letter published in the September–October, 1953, issue of this JOURNAL have further mentioned a point that does affect my old Department, the Geological Survey of India. It is as well to place on record the history of the determination of the Kolhan Series during the period when I was making that determination.

[1] This reply will conclude the discussion. EDITOR.

Spencer and Percival make the new statement that "The onus of re-mapping the area was on the Geological Survey." Geological mapping is wide open to everyone. The Geological Survey of India over its century of history has accepted and published much fine geological work by non-official geologists working for companies, and others. Had Percival and Spencer in "the long years between 1922 and 1938" produced a map of the area with which they were so long associated, I have no doubt that it would not only have cleared up the problem for them and others, but its authority would have been accepted by the Geological Survey of India. However, in the event, the problems would still have been unsettled had the Geological Survey of India not been puzzled by the lack of correlation with adjacent areas and, appreciating the real significance of the conglomerates, re-mapped the area.

The history will be found in the Director's Annual Reports of the Geological Survey of India for the years 1937 and 1938 in Records of the Geological Survey of India, vol. 73, p. 27, and vol. 74, pp. 27–28; also in Memoirs of the Geological Survey of India, vol. LXIII, part 3, pp. 303–308. The area had been mapped by a former senior colleague H. C. Jones in the early 1920's. All familiar with the area, including Jones, Percival and Spencer, were aware of conglomerates higher in the sequence than the supposed basal conglomerate, and were puzzled by them. Gradually I concluded that there may be something wrong with Jones's sequence, partly by the inability to correlate South Singhbhum stratigraphy with adjacent areas. Finally, in 1937, while visiting the iron ore mine at Gua, I paid particular attention to these conglomerates and appreciated that only by re-mapping was the problem of correlation likely to be solved. As a Director of a Geological Survey cannot readily spend time and money on a resurvey of a former senior colleague's work, two of my colleagues made a brief visit with me and we unhesitatingly recommended to the Director that re-mapping was advisable. In making this recommendation, as mentioned in Memoirs LXIII, remarks on these conglomerates by others, including Jones, Spencer, and Percival, were not overlooked but of course the deciding factors were what we had seen in the field and all the problems of lack of correlation with other areas.

The problem was solved and the Kolhan Series established by December 1938, although some work on outlying areas remained. The work was discussed with Percival and Spencer as with others and the revised maps shown to them, as the area was one with which they were closely connected, and the maps removed the former confusion. It is easy to be wise after the event, but it is a sound rule in geology that when in doubt some re-mapping may provide a solution.

I prefer not to discuss further Spencer and Percival's remarks on my views on the origin of the banded hematite quartzite, or jasper. Those interested can find all the evidence leading to my interpretations in my published Memoirs of the Geological Survey of India (including, with plates, examples of preservation of amygdules in re-crystallized lavas!)

My views have not altered since 1941 (ECON. GEOL. 36, p. 366). Evidence of the sedimentary origin of the silica is of a negative kind in that no one

has disproved that the bulk of the silica is sedimentary; on the other hand evidence of secondary silica is positive and widespread. I remain willing to be convinced that the silica was mainly sedimentary. Divorcing the factual description from their interpretation in Spencer and Percival's paper, no facts are presented that are opposed to replacement, but even the facts presented by those authors do indicate that there has been some replacement.

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Dec. 24, 1953

REVIEWS

Pegmatite Deposits of the White Picacho District, Maricopa and Yavapai Counties, Arizona. By RICHARD H. JAHNS. Pp. 105; pls. 24; figs. 5; tbls. 6. University of Arizona, Tucson, 1952. Price, $1.25.

This bulletin represents the first detailed pegmatite investigation in Arizona. Approximately six weeks of field work was done between March 1948 and June 1952.

The White Picacho is one of several pegmatite districts in the Arizona Pegmatite Belt which lies southwest of the Colorado Plateau. This district occurs principally in the Mexican Highland section, but a small part of it is in the Sonoran Desert section of the Basin and Range province. The Arizona Pegmatite Belt is about 250 miles long, 30 to 80 miles wide, and extends southeastward from Lake Mead to points south of Phoenix.

More than 100,000 tons of crude and ground feldspar have been produced in the last three decades in the northern district of the pegmatite belt, a few miles northeast of Kingman. Although known to local miners and prospectors, it was not until 1947 that the pegmatites of the White Picacho district were critically examined and prospected for feldspar. The principal pegmatites were spotted by airplane and these, and other pegmatites, were later prospected and mined on a small scale. Large amounts of high-grade feldspar, spodumene, amblygonite, and other lithium minerals were found in several pegmatites. During 1952 four properties were further explored by diamond-drilling in search for commercial lithium deposits.

As a result of this recent prospecting and exploration of many pegmatites much geologic and economic data became available. These data have been completely and authoritatively evaluated in this bulletin. The geography, general geology, and structural features of the district are described briefly. Three general types of pegmatites are recognized: (1) simple, (2) zoned, and (3) zoned-lithium pegmatites. Variations in form, size, and attitude are noted. Bulges are common in many pegmatites. The bulbous parts of the White Picacho pegmatites are commonly zoned and contain feldspar and lithium minerals in commercial quantities. Most of the large and bulbous pegmatites, containing commercial mineral deposits, are enclosed in foliated varieties of Precambrian schist in contrast to non-commercial deposits in small, thin, and highly irregular pegmatites in granite or other relatively massive rocks.

A description of the internal structure of pegmatites clearly defines the significance of zones, fracture filling, and replacement bodies as a means of appraising these units in relation to their economic potentialities. Each zone is described in detail, and particular references are made to those pegmatites where certain zonal features are best developed. The intermediate zones are the principal source of the lithium minerals; the cores contain most of the feldspar deposits. Although the fracture fillings do not contain commercial deposits of lithium minerals, they serve as guides to lithium mineralization in the enclosing pegmatite.

Seventy-six minerals have been identified in the pegmatites of the district. Five are listed as abundant and widespread; 2 are widespread, or locally abundant; 7 are common, or abundant in at least one pegmatite; 22 are not common; and 40 are rare. These minerals are grouped under four headings: essential, varietal, accessory, and other minerals. In general, mineral descriptions emphasize the economic features. The occurrence of minerals in a particular pegmatite is noted as well as the zone in which they may be located. The essential minerals are microcline, perthite, albite, cleavelandite, and quartz. The varietal minerals include muscovite, biotite, lepidolite, zinnwaldite, spodumene, amblygonite, and tourmaline (schorl and elbaite). The most widespread accessory minerals are apatite, beryl, bismuth minerals, columbite-tantalite, fluorite, garnet, lithiopholite-triphylite, .magnetite, microlite and pyroclore, pyrite and other sulfides, and triplite. Other rarer accessory minerals are allanite, cassiterite, monazite, scheelite and powellite, and spinel.

The sequence of mineral formation is described briefly and summarized in Figure 4, p. 53. A magmatic origin of the pegmatites is proposed and supported by field and laboratory data. Zones are believed to have developed progressively inward from the walls by crystallization of a magma. Fracture fillings and replacement bodies are described and considered to have formed subsequent to pegmatite zones.

Under the heading, Economic Features of the Pegmatite Minerals, feldspar, lithium minerals, beryl, mica, columbite-tantalite, and bismuth minerals and quartz are discussed chiefly according to uses, classification, grade determination, marketing data, and possible economic occurrences and development in the district.

Large lots of feldspar have been mined for testing at the Picacho View, Outpost, Friction, and Midnight Owl mines. The North Morning Star and Midnight Owl mines have produced 26 and 62 tons of amblygonite respectively. Spodumene has been produced from the North Morning Star (20 tons), Lower Jumbo (55 tons), and an unknown amount from the Midnight Owl. Thirteen tons of beryl have been produced at the Midnight Owl mine, 12 tons of bismuthite from the Outpost mine, about 100 tons of muscovite from several mines, and less than one ton of columbite-tantalite from the Midnight Owl mine. The principal mineral commodities of the district are believed to be high-grade feldspar and lithium minerals. Beryl, mica, quartz, and other minerals will be produced most economically as by-products.

In addition to maps and diagrams the bulletin is well illustrated with 15 photographs. A supplement (enclosed loose envelope) contains detailed geologic and economic maps of some of the more promising pegmatites. A complete description of these and other pegmatites is presented in the last section (p. 78–105). The publication of this investigation by one of the outstanding authorities in this country on pegmatites and pegmatite minerals makes available much valuable geologic and economic data which can be tested by mining companies. Companies interested in developing properties for feldspar and lithium minerals in this district should examine pages 57–105 (of this bulletin) for an economic evaluation of deposits. Those interested in the geology and structure of pegmatites will find information on the internal structure and genesis of pegmatites. The amateur mineralogist of Arizona and nearby states can search the bulletin for new localities of many rare and interesting minerals.

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February 5, 1954

Petrography. By HOWELL WILLIAMS, FRANCIS J TURNER AND CHARLES M. GILBERT. Pp. 416; figs. 133. W. H. Freeman and Company, San Francisco, 1954. Price, $6.50.

This book is one of "A Series of Geology Texts, Editors: James Gilluly and W. O. Woodford." The preface states, "This book is concerned more with the description of rocks than with problems of genesis, more with petrography than with the broader field of petrology." It is divided into three parts; Part I, Igneous Rocks, is by Williams; Part II, Metamorphic Rocks, is by Turner; and Part III, Sedimentary Rocks, is by Gilbert. Practically all microdrawings, of which there are many beautiful ones, were made by Professor Williams.

Part I starts off with ten pages dealing with Origin, Magmatic Evolution, Stages of Consolidation of Magma, and Igneous Rock Associations, or Kindreds. Chapter 2 follows on Textures and Classifications of Igneous Rocks. The remaining six chapters deal with the following clans: Gabbro; Alkali Gabbro; Ultramafic and Lamprophyres; Diorite, Monzonite and Syenite; Granodiorite, Adamellite and Granite; and Pyroclastics. One hundred and twenty pages are devoted to the different clans.

Part II consists of five chapters on Metamorphism, its Petrographic Criteria and its Products; Hornfelses and Spotted Slates; Cataclasites, Mylonites and Phyllonites; Slates, Phyllites and Schists of Low Metamorphic Grade; and High Grade Schists, Amphibolites, Granulites and Eclogites. These are covered in 90 pages. The inclusion of eclogites under metamorphic rocks may draw comment. The origin of charnockites is quickly disposed of: "It is better to regard the majority of charnockites as products of deep-seated metamorphism of quartzo-feldspathic rocks many of which were initially igneous" (p. 239).

Part III has six chapters on Origin, Composition and Texture, Sandstones, Argillaceous rocks, Calcareous rocks, and Miscellaneous Sedimentary Rocks. The emphasis given to sedimentary rocks is indicated by the 133 pages devoted to them. This is an advance over most text books since so many geologists today enter the petroleum industry.

The book is well written, well illustrated and nicely printed. The examples of rocks chosen for description are widely distributed and the authors have avoided introducing new rock names. The simplicity of presentation will appeal to students. The book bids fair to becoming the outstanding text book on elementary petrography.

Tafeln zum Bestimmen der Minerale nach äusseren Kennzeichen. By HELLMUT V. PHILIPSBORN. Pp. 244; figs. 290. E. Schweitzerbart'sche Verlagsbuchhandlung, Stuttgart, 1953. Price, DM 17.

A determinative table of minerals can only be judged after much use. Nevertheless, the proved arrangement of earlier editions of the Weisbach-Kolbeck determinative tables promises that there should be no important blunders. Professor von Philipsborn's new edition of the main tables, and his additions of certain auxiliary tables, will be welcomed not only for their updating, but even more for their new features.

The main tables list minerals according to the following scheme: divisions of first rank by luster; of the second, by color (of metallic minerals) or streak (if applicable) or hardness (of nonmetallic minerals with noncharacteristic streak). Some such classes are very large; the average number of minerals per class is a

little over 33. Minerals are serially numbered (1–568), but with duplications (e.g., *casseriterite,* nos. 139, 192, 261, 531), and unnumbered entries (e.g., amesite, clinochlor, delessite, and prochlorite under *penninite,* no. 273). The total number of different minerals included is probably near 500. Some 289 crystal habits are illustrated, both in text and in drawings, in the auxiliary tables on morphology of crystals and aggregates; one chemical table gives tests for the elements; another, minerals arranged by elements; a third, chemical test-results arranged according to the types of test (e.g., flame colors).

The determinative table of refractive indices is arranged in order of n_X-values, with isomorphous series or groups (e.g., olivine) entered together in the position appropriate to the lowest n_X in the whole group (1.635 for olivine, although fayalite has $n_X = 1.835$). The use of n_X may be confusing, for an experimental measurement of n_X may be much more difficult than the measurement of n_Y or n_Z—especially with negative uniaxial minerals such as calcite, and negative biaxial minerals with very perfect cleavage, as muscovite and most other phyllosilicates.

Cross references between the several tables facilitate the finding of data for any mineral. The alphabetical index has about 1400 entries, including obsolete names and synonyms. The inclusion of a glossary of about 250 mineralogical terms in five common European languages (e.g. Probe = sample = échantillon = campione = muestra) is a valuable feature; its arrangement by subject may be debatable.

This outstanding book has come far since Weisbach's first edition (1866); the improvements in size and completeness are indeed notable. If language permitted I would certainly try it in classes.

HORACE WINCHELL

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February 25, 1954

The Charnockite Problem. By C. S. PICHAMUTHU. Pp. 178. Mysore Geologists' Assn., Bangalore, India, 1953. Price, Rs. 8.

Dr. Pichamuthu, as Director of the Mysore Geological Department, has had much opportunity over many years to study personally and to supervise field work on charnockites. This volume reflects his careful review of the problem, and is based upon his Presidential Address before the Annual Meeting of the Mysore Geologists' Association at Bangalore, India. Although the type locality for charnockites is at Pallavaram in Madras, there are more extensive occurrences in Mysore.

Dr. Pichamuthu first considers the nomenclature, followed by the characteristics of the charnockite series, their mode of occurrence, their distribution in India and elsewhere, mineralogy, petrography and economic aspects. He then takes up the knotty question of origin, first summarizing the arguments in favor of an igneous origin and then those in favor of a metamorphic origin—both very objective discussions. Following these excellent summaries, he considers the origin, upheaval of charnockite massifs, history of research outside India, charnockites of Mysore, and then a pointed chapter called "Charnockites and Charnockites" in which he points out that many rocks are called charnockites, which diverge from the original definition. An interesting discussion is his Chapter 15, termed "Discussion." He then considers the geologic history of the charnockite region, which is followed by a summary, and an extensive bibliography of 15 pages, concluded by 5 pages of chemical analyses of charnockites.

A valuable chapter is his "Chapter 12, History of Research Outside of India," in which he gives abstracts, chronologically, of practically all papers that have appeared throughout the world on the subject of charnockites.

Dr. Pichamuthu does not attempt to settle the problem of charnockites. Rather, he has given a masterly review of the various facets of the problem and has brought together in one book for the first time the existing knowledge relating to charnockites. Every student of petrology interested in this type of rock and the problems it presents will wish to refer to this book. While Dr. Pichamuthu does not commit himself as to which origin he thinks the evidence supports most strongly, the reader is left with a slight impression that he inclines toward an igneous origin. The reviewer's own brief observations of charnockite areas in India also tend toward an igneous origin.

Petrologists the world over owe thanks to Dr. Pichamuthu for preparing this valuable book and also to Mr. P. S. Narayan of Bangalore who made its publication possible.

ALAN M. BATEMAN

NEW HAVEN, CONN.,
March 8, 1954

World Population and Production. By W. S. WOYTINSKY and E. S. WOYTINSKY. Pp. 1268; tbls. 338; pls. 338. Twentieth Century Fund, New York, 1953. Price, $12.00.

This husky volume was prepared solely by the two widely travelled authors with the financing of the Rockefeller Foundation and the Twentieth Century Fund. Its purpose is "to outline world economic forces and trends in our present fateful era, in which modern mechanized civilization is spreading over the world and becoming universal. . . . The authors seek to present a picture of the entire globe, of which every country and everyone of us humans is a part."

There are five parts: Part I, Man and His Environment; Part II, World Needs and Resources; Part III, Agriculture; Part IV, Energy and Mining; Part V, Manufactures.

Under Part I are chapters on the earth, peoples, migration, cities, births and deaths, marriages and divorces, health, and the future of world population. Part II deals with consumer needs and outlays, consumption, standards of living, resources, abundance and scarcity, and economic patterns. Part III deals with all the various phases of agriculture. Part IV, on Energy and Mining, has chapters on mining and world economy, minerals, coal, petroleum and gas, and the economics of energy and power. Part V covers all the chief phases of manufacturing.

Part IV, of greatest interest to most of the readers of this Journal gives a well balanced picture of mineral resources, accompanied by statistical data. It gives their uses, distribution, something about reserves, production, and other data. It is a compact assemblage of data on world wide mineral resources, taken up mineral by mineral.

The book is a monumental piece of compilation and one wonders how two authors could possibly have made such a survey by themselves.. It is a book that every library should have and will prove a valuable reference to all geographers and other scientific investigators.

BOOKS RECEIVED

FRANK G. LESURE

U. S. Geological Survey—Washington, D. C., 1954

Prof. Paper 230. Geology of the Elkton Area, Virginia. PHILIP B. KING. Pp. 82; pls. 6; figs. 19; tbls. 18. Price, $1.75. *Detailed report on area containing interesting problems of stratigraphy, structure, geomorphology and economic geology.*

Prof. Paper 235. Geology and Ore Deposits of the West Slope of the Mosquito Range. CHARLES H. BEHRE, JR. Pp. 176; pls. 20; figs. 101.

Prof. Paper 242. Larger Invertebrate Fossils of the Woodbine Formation (Cenomanian) of Texas. LLOYD WILLIAM STEPHENSON. Pp. 226; pls. 59; figs. 8; tbl. 1. *Description of 140 genera including 281 species and varieties.*

Bull. 988-G. Geology of the Area Adjacent to the Free Enterprise Mine, Jefferson County, Montana. W. A. ROBERTS AND A. J. GUDE, 3D. Pp. 143–155; pls. 3; figs. 2.

Bull. 994. Structures and Forms of Basaltic Rocks in Hawaii. CHESTER K. WENTWORTH and GORDON A. MACDONALD. Pp. 98; figs. 45. Price, 45 cents. *Description of physical features of basaltic rocks and their origin.*

Bull. 996-B. Hawaiian Volcanoes During 1950. R. H. FINCH AND GORDON A. MACDONALD. Pp. 27–98; pls. 2; figs. 20. Price, 55 cents. *Report of Hawaiian Volcano Observatory.*

Bull. 998-C. Some Lead-Zinc and Zinc-Copper Deposits of the Ketchikan and Wales Districts, Alaska. G. D. ROBINSON and W. S. TWENHOFEL. Pp. 59–84; pls. 2; figs. 6. Price, 55 cents. *Description of deposits at Moth Bay, Mahoney Creek and Dora Lake.*

Bull. 1002-C. Geophysical Abstracts 154, July–September 1953. MARY C. RABBITT, S. T. VESSELOWSKY and others. Pp. 123–178. Price, 25 cents. *Abstract numbers 14599–14804.*

Circ. 316. Reconnaissance for Radioactive Deposits in the Eagle-Nation Area, East-Central Alaska, 1948. HELMUTH WEDOW, JR. Pp. 9; pl. 1; fig. 1; tbls. 3.

Circ. 317. Reconnaissance for Radioactive Deposits in the Manley Hot Springs–Rampart District, East-Central Alaska, 1948. ROBERT M. MOXHAM. Pp. 6; figs. 2; tbl. 1.

Circ. 321. Occurrences of Uranium-Bearing Minerals in the St. Kevin District, Lake County, Colorado. C. T. PIERSON and Q. F. SINGEWALD. Pp. 17; figs. 8; tbl. 1.

Circ. 322. The East Slope No. 2 Uranium Prospect, Piute County, Utah. DONALD G. WYANT. Pp. 6; pl. 1; figs. 2; tbls. 2.

Circ. 324. Stratigraphic Sections of the Phosphoria Formation in Wyoming, 1951. T. M. CHENEY, R. P. SHELDON, R. G. WARING and M. A. WARNER. Pp. 22; figs. 2.

Circ. 325. Stratigraphic Sections of the Phosphoria Formation in Wyoming, 1952. R. P. SHELDON, E. R. CRESSMAN, L. D. CARSWELL and R. A. SMART. Pp. 24; figs. 2.

Circ. 326. Stratigraphic Sections of the Phosphoria Formation in Montana, 1951. J. A. PETERSON, R. F. GOSMAN and R. W. SWANSON. Pp. 27; figs. 2.

Circ. 327. Stratigraphic Sections of the Phosphoria Formation in Idaho, 1950–51. R. A. SMART, R. G. WARING, T. M. CHENEY and R. P. SHELDON. Pp. 22; figs. 2.

Circ. 328. Reconnaissance for Radioactive Deposits in the Lower Yukon–Kuskokwim Region, Alaska, 1952. WALTER S. WEST. Pp. 10; pls. 2; figs. 2; tbls. 4.

U. S. Atomic Energy Commission—Oak Ridge, Tennessee, 1954.

Fifteenth Semiannual Report. Pp. 151.

RME-3069. Report of Geological Reconnaissance in South-Central Montant and Northwestern Wyoming. MARCELLUS H. STOW. Pp. 34; figs. 3.

RME-4028. Prospecting with a Counter. ROBERT J. WRIGHT. Pp. 70; figs. 10. Price, 25 cents. *Summarizes information on field counters; their operation, use, abuse; application to prospecting, mining and geologic problems. Emphasizes practical problems.*

Geology of the Golden-Green Mountain Area, Jefferson County, Colorado. STANLEY O. REICHERT. Pp. 96; pls. 13; figs. 49; tbls. 5. Quarterly of the Colorado School of Mines, Golden, 1954. Price, $1.50. *Detailed study of stratigraphy of area.*

An Analysis of Selected Indiana Coals by the Particle Count Method. RANARD JACKSON PICKERING. Pp. 23; pl. 1; figs. 4. Indiana Geological Survey Progress Rept. 6, Bloomington, 1954. Price, 50 cents. *Description of method and suggested commercial and geological applications.*

Directory of Missouri Geologists. Compiled by OLIVER RUDOLPH GRAWE. Pp. 15. Bull. Missouri School of Mines and Metallurgy, Rolla, 1954.

North Dakota Geological Survey—Grand Forks, 1954.

Rept. of Inv. 14. Surface Structure of Western Stark County and Adjacent Areas of North Dakota. JOHN W. CALDWELL. 1 sheet.

Rept. of Inv. 15. Structural Geology of the Skarr–Trotters Area, McKenzie and Golden Valley Counties, North Dakota. STANLEY P. FISHER. 1 sheet.

Rules and Regulations for the Conservation of Oil and Natural Gas and Appendix. Pp. 33. Oregon Dept. of Geology and Mineral Industries, Misc. Paper 4, Portland, 1954. Price, 50 cents.

Sixtieth Annual Rept., Vol. LX, Pt. IX, 1951. Pp. 56; pls. 8; fig. 1. Ontario Dept. of Mines, Toronto, 1953. *General geology, description of gold claims.*

Acta Cuyana de Ingeniera—Universidad Nacional de Cuyo, San Juan, Argentina, 1953.

Vol. 1, No. 1. Nuevo Metodo Sismografico para el Estudio de Fundaciones de Diques de Embalse. FERNANDO VOLPONI. Pp. 34; figs. 15. *Descriptions of seismograph method applied to exploration at shallow depths, especially in river beds.*

Vol. 1, No. 2. Un Metodo Expeditivo de Compensacion. BERNHARD H. DAWSON. Pp. 37–48.

Vol. 1, No. 3. Estudio Experimental del Metodo de Aforo por Titulacion. ENZO OSCAR MACAGNO and LUIS ROVIRA. Pp. 32; figs. 4; tbls. 6.

Vol. 1, No. 4. Analisis Cualitativo del Movimento de los Liquidos. ENZO OSCAR MACAGNO. Pp. 36; figs. 32.

L'Hydrogéologie des Roches Calcareuses. IVAN DE RADZITZKY D'OSTROWICK. Pp. 199; figs. 59. Extrait du Bulletin de la Société Royale d'Études Géologiques et Archéologiques, "Les Chercheurs de la Wallonie," t. XV, Dinant, Belgique, 1953. Price, 160 frs belges. *Describes formation of caves and solution features, mineral deposits in caves and importance for ground water resources.*

Il Deposito di Bauxite di S. Giovanni Rotondo. ANTONIO CAVINATO. Pp. 497–503; figs. 4. Industria Mineraria, fasc. of November, 1953, Anno IV, Faenza, 1953.

Bull. of the Geological Survey of Japan, Vol. 4, No. 9. Pp. 54. Geol. Survey of Japan, Hisamoto-cho, Kawasaki-shi, 1953. *Papers on water resources; land creep; natural gas; iron sulfide ore in Hokkaido.*

Geologiska Föreningens, Band 75, Häfte 4. Pp. 437–533. Geol. Soc. of Stockholm, 1953. *Includes a short review of the geologic literature of Sweden in 1952 with abstracts in German, French or English.*

Geological Survey of Tanganyika—Dar es Salaam, 1953.

Geologic Map of Mpwapwa-South B37/M1. Revised by B. N. TEMPERLEY. Scale 1: 125,000.

Short Paper No. 28. The Geology of the Mhukuru Coalfield. D. A. HARKIN. Pp. 66; pls. 6. Price, Shs. 7/50/.

SCIENTIFIC NOTES AND NEWS

PARKER D. TRASK has recently returned from Egypt where he made a survey of that country's mineral resources for the Foreign Operations Administration and the U. S. Geological Survey.

THE SOCIETY OF EXPLORATION GEOPHYSICISTS announces the election of the following officers for 1954–1955: PAUL L. LYONS, Tulsa, president; ROY F. BENNETT, Oklahoma City, vice-president; HUGH M. THRALLS, Tulsa, secretary-treasurer.

THE 1954 PACIFIC COAST REGIONAL CONFERENCE ON CLAYS AND CLAY TECHNOLOGY is to be held at the University of California, Berkeley, on June 25 and 26. The subject of the Conference will be the clay-water system. Papers related to permeability of liquids through formations, swelling of clays, viscosity and plasticity will be presented by workers in the fields of petroleum, soil science, soil mechanics, geophysics, and ceramics. The Clay Minerals Committee of the National Research Council is a co-sponsor of the Conference.

C. W. PEGG, JR., geologist for Ownamin Limited, has left British Columbia for Bathurst, N. B.

DONALD BOURNE, formerly chief geologist for the consulting engineering firm of Hill, Legg and Hemsworth, Vancouver, is now geologist for Eldorado Mining and Refining, Ltd.

ALFRED W. G. WILSON, for many years Chief of the Mineral Resources Division of the Federal Department of Mines, and at the time of his retirement Chief Technical Consultant, died in Ottawa on January 9th.

W. G. PYNE MERCIER has been appointed Government Mining Engineer of the Union of South Africa.

K. C. HEALD announces he has set up as petroleum consultant at 605 Continental Life Bldg., Fort Worth 2, Texas.

J. K. MACDONALD has been transferred from African Chrome Mines, Ltd., to the Selukwe Peak Mine of Rhodesia Chrome Mines, Ltd.

JOHN BRACKEN, until recently with the O'OKiep Copper Company, Ltd., Nababeep, S. Africa, is now on the staff of Dominion Wabana Ore, Ltd., Wabana, Nfld.

K. A. PHILLIPS has taken a position in the Geological Survey Department, Lusaka, Northern Rhodesia.

HORACE J. FRASER, general manager of Falconbridge Nickel Mines and former professor of geology at the California Institute of Technology, has been elected president of the Canadian Metal Mining Association.

WILLIAM J. MCCAUGHEY, Professor Emeritus of mineralogy, Ohio State University, has been selected as recipient of the Bleininger medal. The presentation was made on March 12 at a meeting of the Pittsburgh section of the American Ceramic Society.

A conference on Latin American geology was held March 29–30, at the University of Texas, Austin, organized by the DEPARTMENT OF GEOLOGY and the INSTITUTE OF LATIN AMERICAN STUDIES of the UNIVERSITY OF TEXAS to bring together workers in the field of Latin American geology in order to survey the present state of research and to focus attention on the areas in which more investigation is needed. Nine papers were presented and discussed.

C. KIETH BELL has been appointed to the position of geologist, grade 2, with the Geological Survey of Canada, Department of Mines and Technical Surveys.

THE NATIONAL SCIENCE FOUNDATION sponsored a Conference on Bioluminescence at Pacific Grove, California, March 29–April 1, 1954, held under the auspices of the Committee on Photobiology, National Research Council. Bioluminescence, or "cold light" is the name given to the emission of light without heat by certain animals and plants.

Senators MURRAY and MANSFIELD introduced a bill, S. 2432, into the Senate of the United States to create a "United States Department of Mineral Resources" as an executive department to be headed by a Secretary of Mineral Resources, an Undersecretary and four Assistant Secretaries. It is proposed to transfer to this Department, from the Department of the Interior, the Geological Survey, the Bureau of Mines, Divisions of Minerals and Fuels and of Oil and Gas, Defense Minerals Exploration Administration, Defense Solid Fuels Administration, Petroleum Administration for Defense, Defense Material Procurement Agency, and such portions of the Bureau of Land Management relating to mineral patents and leases including coal, oil, gas, and nonmetallic minerals.

MORRIS M. LEIGHTON will retire as Chief of the Illinois State Geological Survey after thirty-one years of service on July 1, 1954, and will be succeeded by Dr. JOHN C. FRYE, now State Geologist and Professor of Geology, University of Kansas. During the ten years that Dr. Frye has administered the Kansas Geological Survey he has made it one of the better surveys of the country. Dr. Leighton is retiring to complete his research on the Pleistocene geology of Illinois which he began thirty-five years ago. He will continue as Business Editor of this Journal.

The Cleveland-Cliffs Iron Company in Ishpeming, Michigan, has announced the appointment of BURTON H. BOYUM as chief geologist.

R. H. B. JONES, consulting geologist for the Oliver Iron Mining Division of the U. S. Steel Corporation, has been named staff geologist for U. S. Steel's Columbia-Geneva Steel Division.

WALTER J. WILLIAMS has left the Atomic Energy Commission to become vice-president of the Taconite Contracting Corporation, a Lake Superior subsidiary of the Erie Mining Company. Mr. Williams, called a "key man" in atomic energy development of the last decade by AEC chairman Lewis L. Strauss, has been general deputy manager of the nuclear energy program since 1951. Previous to that time he was manager of field operations for the AEC. His new post will include development of taconite in the Lake Superior iron ore region.

H. R. COOKE, head of a geological exploration project for the Northern Peru Mining and Smelting Company in Peru, has completed his contract and is now on vacation in Europe. He expects to return to the United States in June.

EVERETT GRAFF, mining geologist for the Cerro de Pasco Mining Company, has returned to Peru after a brief visit in the United States. While in Spokane, Washington, Mr. Graff, who has his own consulting business in Lima, expressed an optimistic view of mining opportunities in Peru.

A. G. PENTLAND, geologist for Sheep Creek Gold, Ltd., has been elected president of West Kootenay Mine Safety Association at Nelson, British Columbia.

G. C. MONTURE, chief of the mineral resources division of the Canadian Department of Mines and Technical Surveys, is acting as mining advisor for a World Bank survey group in Malaya.